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# THE JOURNAL OF GEOLOGY

VOLUME 57 NOVEMBER 1949 NUMBER 6

PETROGENESIS OF PERTHITE PEGMATITES IN THE BLACK HILLS, SOUTH DAKOTA
Riad A. Higazy 555
SANDSTONE TYPES: THEIR ABUNDANCE AND CEMENTING AGENTS
Sefton L. Tallman 582
RECENT RESULTS OF INVESTIGATIONS ON THE FELDSPARS . Alexander Köhler 592
DIE HÄUFIGKEIT DER EDELGASE AUF DER ERDE UND IM KOSMOS . Hans E. Suess 600
A NOTE ON THE GEOCHEMISTRY OF THALLIUM . . . . . Kalervo Rankama 608
GEOLOGICAL NOTES
Trivoli Sandstone of Williamson County, Illinois . . . . . Raymond Siever 614
A Proposed Method for the Measurement of Geologic Temperatures . . I. I. Friedman 618
REVIEWS
*Cretaceous and Tertiary Subsurface Geology: The Stratigraphy, Paleontology, and Sedimentology of Three Deep Test Wells on the Eastern Shore of Maryland* (W. C. Krumbein) . . . . . 620
"Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit," Hans Poser (Ernst Antevs) . . . . . 621
*Soil Mechanics in Engineering Practice*, Karl Terzaghi and Ralph B. Peck (Parker D. Trask) . 622
*Physical Geology*, Chester R. Longwell, Adolph Knopf, and Richard F. Flint (Lou Williams Page) . . . . . 623
*The Religion of the Modern Scientist (Neo-materialism)*, S. W. Tromp (Kirtley F. Mather) . . . 625
*Mining Geology*, H. E. McKinstry, with sections by Stanley A. Tyler, E. N. Pennebaker, and Kenyon E. Richard (Olaf N. Rove) . . . . . 625
*The Iron Resources of California* (Charles F. Park, Jr.) . . . . . 626
COMMUNICATIONS AND ANNOUNCEMENTS . . . . . 628
ERRATA . . . . . 628
INDEX . . . . . 629

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"Ripple Mark in Rock Salt of the Salina Formation." By DALE W. KAUFMAN and C. B. SLAWSON

"Quartz Crystals with Clay and Fluid Inclusions." By STEPHEN TABER

"Stereoscopic Projection for Demonstration in Geology, Geomorphology, and Other Natural Sciences." By PH. H. KUENEN

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REFERENCES—Under the heading "References Cited" at the end of the paper should be listed all literature cited, arranged alphabetically by authors and chronologically under each author, using the following form:

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——— (1884*b*) Guidebook for excursions in the Big Ben Mountains, New York, Macmillan Co.

SMITH, A. V. (1926) Discovery of fish remains in Ordovician rocks of the Black Hills (abstr.): Geol. Soc. America Bull. 52, p. 27.

——— and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76, 3 pls.

——— — ——— (1929*b*) Revision of Ordovician stratigraphy of the Black Hills: Jour. Geology, vol. 37, pp. 680–695.

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VOLUME 57 NUMBER 6

# THE JOURNAL OF GEOLOGY

*November 1949*

## PETROGENESIS OF PERTHITE PEGMATITES IN THE BLACK HILLS, SOUTH DAKOTA[1]

RIAD A. HIGAZY[2]
Farouk I University, Alexandria, Egypt

ABSTRACT

The study of some pegmatites of the Black Hills, South Dakota, has led to the conclusion that metasomatic replacement played a significant role in their formation. Microcline microperthites of these pegmatites were not products of magmatic crystallization but were formed by complex processes of perthitization. The hypothesis of the metasomatic origin of the perthites is supported by chemical analyses and microscopic features of the perthites and country rocks and by field observations.

### STATEMENT OF PROBLEM

The origin of the potash-rich rocks, whether deep seated or extrusive, is still an unsolved problem. Some authors explain them as products of magmatic crystallization, whereas others believe in a secondary mode of origin of the potash. Final conclusions regarding the genesis of these rocks can be drawn only on the basis of field and laboratory investigations of individual occurrences.

The author studied some potash-rich pegmatites in T. 2 S., R. 6 E., near Glendale, Pennington County, South Dakota (fig. 1). Excellent outcrops and mine cuts facilitated the study. The pegmatite mines studied include the Glendale, in the northernmost portion of NW. ¼ Sec. 23, T. 2 S., R. 6 E.; the Soda Spar, in the SE. ¼ Sec. 22, T. 2 S., R. 6 E.; the Johnson Dike, in the SE. ¼ NW. ¼ Sec. 22, T. 2 S., R. 6 E.; and several smaller unnamed bodies. Several pegmatites, such as the Three Falls Lode in the SW. ¼ Sec. 22, T. 2 S., R. 6 E., previously mined, were also examined.

Smaller pegmatites are very numerous in this area. The map of pegmatites of the Beecher Rock Basin, T. 4 S., R. 4 E., in Custer County, South Dakota, made by Gwynne (1944) shows clearly their wide distribution.

Some of the largest pegmatites of the Black Hills in the neighborhood of both Custer and Keystone were also visited. These were the Buster, 2½ miles south-southwest of Custer, in the SW. ¼ Sec. 2, T. 4 S., R. 4 E.; the New York, in the NE. ¼ Sec. 18, T. 4 S., R. 4 E.; the Peerless, in the NE. ¼ SE. ¼ Sec. 8, T. 2 S., R. 6 E.; the Hugo, in the southernmost portion of SE. ¼ Sec. 8, T. 2 S., R. 6 E.; the White Cap, in the S. ½ Sec. 9, T. 2 S., R. 6 E.; and the Etta, in the NW. ¼ NW. ¼ Sec. 16, T. 2 S., R. 6 E.

The structure of most of these pegma-

[1] Manuscript received October 8, 1948.

[2] On a study leave at the University of Chicago.

tites has been described by other writers: Apsouri (1940), Connolly and O'Harra (1929), Guiteras (1940), Fisher (1942 and 1945), Landes (1928), Lincoln (1927), Page and Norton (1946), and Tullis (1939). Few chemical or microscopic data have been published. The author, therefore, analyzed some perthites and country rocks and examined 65 slides of these rock types. Still more data are needed to make certain the conclusions drawn.

## GEOLOGY OF PEGMATITE AREA

The general geology of the Black Hills has been discussed by Darton and Paige (1925). The Black Hills uplift, an outlier of the Rocky Mountains, is an elliptical area whose long axis trends northwest. The core is composed of metamorphic and eruptive rocks surrounded by concentric outcrops of Paleozoic and Mesozoic sediments. The pre-Cambrian central portion and the outwardly dipping sediments are overlain by Oligocene for-

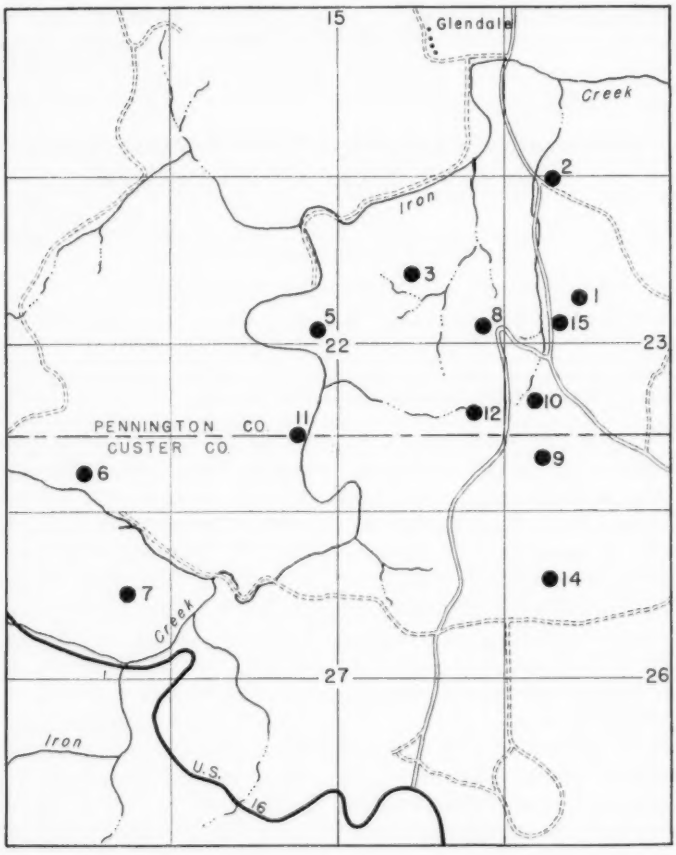



FIG. 1.—Glendale area, showing location of perthites

nations. Hypabyssal dikes and sills of Tertiary age intrude the pre-Cambrian metamorphic rocks and the Paleozoic and Mesozoic sediments (Kirby, 1932).

The pre-Cambrian rocks are metamorphosed sediments and basic igneous rocks, granites, and granite pegmatites. The sediments are conglomerate, quartzite, slate, phyllite, mica schist, garnet and staurolite schists, and marbles (Van Hise, 1890). The basic rocks, originally dioritic or gabbroic, are amphibolites and hornblende schists. In places metamorphism makes it difficult to distinguish between metabasics and metasediments (Connolly, 1927).

The pegmatite granite of Harney Peak is a coarse-grained potassic granite composed principally of microcline-perthite, quartz, and muscovite with an unusually large amount of tourmaline. The pegmatites are also rich in potash and consist chiefly of those minerals which form the pegmatitic granite.

Intense metamorphism and lack of fossils (Connolly, 1927) make it difficult to determine the age of the sediments. Most investigators assign these rocks to the Algonkian. Davis (1926) determined the age of the uraninite from the pegmatites of the southern Black Hills as 1,167 million years. The corrected age, applying Wickman's graphic chart (1943) of $U/(U + Pb)$ and $U/(U + Th)$, is 1,450 million years. The Black Hills pegmatites are, therefore, Archean.

## FIELD RELATION OF PEGMATITE AND ASSOCIATED SCHISTS

Recent studies of the Black Hills pegmatites show that most of the mined pegmatites are zoned (Page and Norton, 1946). The zones occur in a definite sequence, and, although any number of different zones may be missing, the sequence of the existing zones will remain the same. Perthite may be either an essential constituent of the zones or may form one of the zones. The Hugo pegmatite and the High Climb Pit, 6 miles north of Custer (Fisher, 1942, pl. 19 and fig. 4) are examples of the first type. Perthite is the principal constituent of some of the different zones of the Soda Spar pegmatite, the Johnson Dike, and the Three Falls Lode.[3]

Examples in which perthite alone forms one of the zones are the White Cap and the John Ross Lode, 4 miles west of Custer (Fisher, 1945, pp. 80–82). The Glendale pegmatite consists of perthite only. Fisher (1945, p. 10) has shown that pegmatitic masses composed almost entirely of perthite are common.

Abundant pegmatitic pockets in numerous unzoned bodies in many cases are composed almost entirely of perthite, commonly associated with albitites.

Perthite is also found as metacrysts in the perthitized schists and as coarse phenocrysts in the granites of the area. The unaltered schist is composed essentially of muscovite, biotite, and quartz with minor amounts of garnet. The schist in places is feldspathized and/or tourmalinized.

The feldspathized schists contain from 25 to 80 per cent of feldspar. The feldspar may be a perthite, an albite, or a microcline.

An outcrop of the perthite schist (pl. 1, *A*, *D*) about 36 yards long, lies in the NE. $\frac{1}{4}$ Sec. 28, T. 2 S., R. 6 E. The foliation of the perthitic schist, of the schist inclusions commonly found in the pegmatites, and of the isolated schist outcrops all have the same trend. The perthite of the schist appears mainly along the schistosity as small ovoid masses less than 1 mm. in diameter, grading into masses more than 3 cm. long, which tend

[3] For location see p. 555 and fig. 1.

to assume the rectangular outlines of a true feldspar (pl. 1, *A*). Some of the larger metacrysts have biotite laminae curved around them, indicating that the growing feldspar crowded aside the folia of the schist. The ovoid masses sometimes join and form a pinch-and-swell structure. Perthites crowded with many megascopic schist inclusions grade into uncontaminated perthite (pl. 1, *B* and *C*). Replacement rather than viscous intrusion and magmatic crystallization seems to be the best explanation of these phenomena.

Microscopic features support this view. The inclusions in the perthite metacrysts are brown biotite and quartz. The biotite has the same optical properties as the schist. The quartz inclusions are more or less rounded and elongated in a direction parallel to that of the schistosity of the matrix. Some of them still

PLATE 1

*A*, Perthite of the perthite schist assuming the rectangular outline of the feldspars (white with dark schist inclusions). Pushing-aside of the schist foliae is best shown at the corners of the crystal *P*.

*B*, Perthite, uncontaminated crystal (*white*); another smaller crystal appears to the right of the figure. The groundmass is schist relic, showing black quartz and lighter-colored micas.

*C*, A crystal of perthite of the perthite schist, showing inclusions of the schist in a megascopic sieve texture. Parts of the crystal are uncontaminated (*white*). Inclusions are dark.

*D*, Perthite schist, showing many perthite crystals (white with light-gray schist inclusions). Schist relic is dark and forms the groundmass.

PLATE 2

*A*, Perthitic schist showing albite (*A*) and string perthite (*P*) with faint grating. Note (1) rounded quartz inclusions (schist relic) in both albite and perthite, (2) the cementing material between some quartz grains is still preserved (*C*), (3) the trend of the quartz inclusions is roughly northeast-southwest in the figure, and (4) biotite inclusions (*B*). 12×.

*B*, Feldspathized schist showing albite (*A*)—polysynthetic twinning—and potash feldspar (*PF*)—simple twinning, Carlsbad law. Notice (1) rounded inclusions of quartz in the feldspars (in some feldspar crystals the inclusions appear white, in others they are extinguished); (2) the trend of inclusions is the same in the different feldspar crystals—northeast-southwest; and (3) the pushing-aside of the biotite flakes of the original schist (*B*) at the corners of the feldspars. 12×.

*C*, Albitite. Notice the cementing material (*C*) between some quartz grains. Quartz (*Q*), albite (*A*), tourmaline (*T*). The albite crystals show polysynthetic twinning, all trending in a roughly parallel direction. Most of them do not show up clearly in the figure. 12×.

*D*, Section of a perthite parallel to (001), showing patch (*white*) and microcline (*grated*). Note the microcline island in the albite patch in optical continuity with host microcline. Note also the extinguished ungrated portion of the microcline. Albite and microcline edges show embayments. 41×.

*E*, Section of a perthite parallel to (010), showing albite veins and patches (*white*) with numerous microcline islands (*gray*). 12×.

*F*, Section of a perthite parallel to (001), showing patch albite (polysynthetic twinning) and microcline (*grated*). Note microcline relics in the patch, showing optical continuity with the host microcline, and irregular resorbed contacts. 12×.

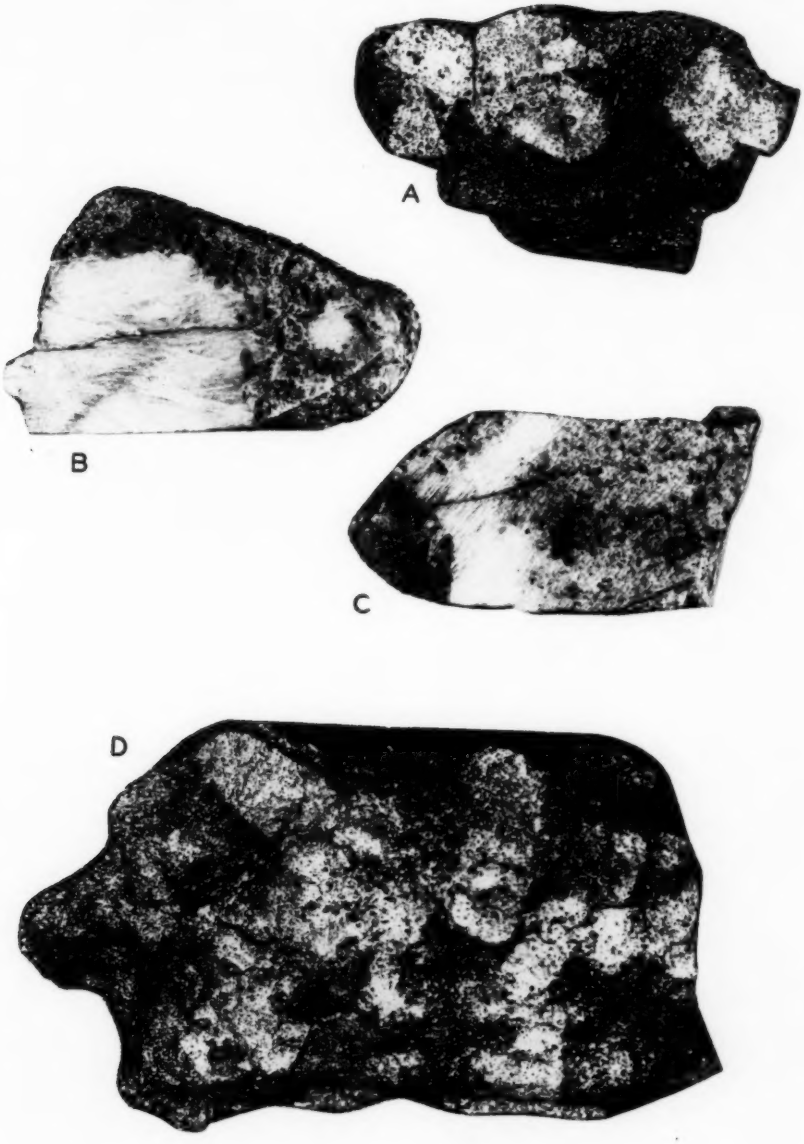
PLATE 3

*A*, Patch and vein perthite (no. 13, table 1), section parallel to (010), albite (*white*), microcline (*gray*). Note (1) fine strings and thin films between the veins, (2) grading of the strings and films into veins near the center, (3) grading of the veins into patches, and (4) microcline relics in the patch at the lower portion. 12×.

*B*, Film, vein, and patch perthite (no. 5, table 1), section parallel to (010), albite (*white*), microcline (*dark gray and black*). Note (1) abundance of thin films, (2) grading of the films into veins, (3) complex texture of the veins grading into patches, and (4) microcline islands in the patch near the center. 12×.

*C*, Film and vein perthite (no. 7, table 1), section parallel to (001), albite (*black*), microcline (*finely grated*). Note abundance of regularly elongated veins and thin films. The latter grade into the former. 12×.

*D*, Vein perthite (no. 10, table 1), section parallel to (010), albite (*white*), microcline (*grated*). Note the roughly round cross section of veins. 12×.

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| Oxides | 1 |
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| $SiO_2$ | 63. |
| $TiO_2$ | 0. |
| $Al_2O_3$ | 21. |
| $Fe_2O_3$ | 0. |
| MgO | 0. |
| CaO | 0. |
| BaO | 0. |
| $Na_2O$ | 2. |
| $K_2O$ | 11. |
| $H_2O$ | 0. |
| Total | 100. |

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ave the yellowish-green chloritic or the ellowish-brown ferriferous cementing aterial, like that of the quartz of the riginal unaltered schist (pl. 2, *A*). The lbite of the schist has formed in the ame manner as the perthite. Plate 2, *A* hows inclusions of rounded quartz of the chist in the albite. They are elongated in he direction of the schistosity. Albite rom a schist in the NE. ¼ Sec. 8, T. 2 S., . 6 E. (across the east-west road from he mill of Consolidated Feldspar Cor-

## COMPOSITION OF PERTHITES AND RELATED ROCKS

Fifteen perthites, one cleavelandite, and three samples of the country rocks were analyzed chemically by the writer at the University of Chicago. Specimens of perthite and microcline from the Hugo pegmatite, obtained from Dr. L. Page, of the U.S. Geological Survey, were analyzed by Mr. B. Bruun. The analyses of the sixteen perthites and the microcline appear in table 1, the analysis of the

TABLE 1

ANALYSES OF PERTHITES*

| Oxides | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | Average | 16 | 17 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $O_3$..... | 63.89 | 65.13 | 64.29 | 63.67 | 64.48 | 64.36 | 63.96 | 64.13 | 64.11 | 64.30 | 64.29 | 64.48 | 65.10 | 64.97 | 65.07 | 64.41 | 64.69 | 64.20 |
| $O_3$..... | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| $_2O_3$..... | 21.61 | 20.70 | 19.40 | 20.08 | 19.54 | 19.53 | 19.39 | 18.92 | 19.62 | 19.37 | 19.76 | 19.66 | 19.55 | 19.41 | 19.04 | 19.72 | 19.79 | 19.38 |
| $_2O_3$..... | 0.09 | 0.09 | 0.09 | 0.07 | 0.09 | 0.09 | 0.13 | 0.06 | 0.06 | 0.09 | 0.09 | 0.12 | 0.10 | 0.07 | 0.13 | 0.09 | 0.17 | 0.05 |
| gO..... | 0.00 | 0.03 | 0.11 | 0.02 | 0.00 | 0.05 | 0.05 | 0.07 | 0.12 | 0.03 | 0.06 | 0.06 | 0.00 | 0.02 | 0.03 | 0.04 | 0.01 | 0.02 |
| aO..... | 0.78 | 0.25 | 0.32 | 0.49 | 0.79 | 0.82 | 0.46 | 0.15 | 0.37 | 0.27 | 0.27 | 0.21 | 0.07 | 0.26 | 0.17 | 0.38 | 0.02 | 0.02 |
| nO..... | 0.02 | 0.05 | 0.06 | 0.02 | 0.01 | 0.02 | 0.17 | 0.03 | 0.04 | 0.05 | 0.03 | 0.05 | 0.02 | 0.05 | 0.05 | 0.04 | n.d. | n.d. |
| $_2O$.... | 2.72 | 2.94 | 0.38 | 1.96 | 2.74 | 2.96 | 2.80 | 2.12 | 2.55 | 1.43 | 2.89 | 2.12 | 2.98 | 3.06 | 3.29 | 2.46 | 3.19 | 1.21 |
| O..... | 11.08 | 10.58 | 15.62 | 13.68 | 12.52 | 11.94 | 12.41 | 14.03 | 12.84 | 14.02 | 13.30 | 13.00 | 11.58 | 12.35 | 11.94 | 12.66 | 11.95 | 14.95 |
| O..... | 0.31 | 0.17 | 0.16 | 0.18 | 0.21 | 0.29 | 0.36 | 0.25 | 0.22 | 0.23 | 0.21 | 0.28 | 0.16 | 0.12 | 0.16 | 0.22 | 0.27 | 0.27 |
| Total.. | 100.50 | 99.94 | 100.43 | 100.17 | 100.38 | 100.06 | 99.73 | 99.76 | 99.93 | 99.79 | 99.90 | 99.98 | 99.56 | 100.31 | 99.88 | 100.02 | 100.00 | 100.10 |

* 1–15: Perthites (see fig. 1 for location); analyst, R. A. Higazy. 16: Perthite (Hugo pegmatite); analyst, B. Bruun. 17: Microne (Hugo pegmatite); analyst, B. Bruun.

oration near Keystone) contains abunant quartz inclusions which clearly ow a sieve texture (pl. 2, *B*). All the clusions in the feldspar crystals are ongated in a northeast-southwest diection in agreement with that of the histosity and the quartz of the matrix. he albite schist in some cases grades to albitites composed essentially of alite, quartz, and micas. In these albitites ie cementing quartz material can still e observed between some quartz grains, nd the albite is generally elongated in ie direction of the schistosity (pl. 2, *C*). hese transitional rocks, containing me relic features of the schist, suggest metasomatic origin for both the rthite schists and the albitites.

cleavelandite is given in table 2, and the analyses of the country rocks are shown in table 3.

The normative potassium feldspar (Or), sodium feldspar (Ab), and lime feldspar (An) of each of the analyzed perthites, microcline, and cleavelandite were calculated, using the equivalent molecular percentage instead of the molecular percentage (Niggli, 1936). This is shown in table 4. The Or is the molecule $KAlSi_3O_8$, composed of one atom of potassium, one of aluminum, and three of silicon; the Ab is the molecule $NaAlSi_3O_8$; and the An is $CaAl_2Si_2O_8$. The percentages of the three feldspar molecules do not differ much from those obtained with the weight percentages in calculat-

ing the norm. The sum of the three molecules forms in most cases 98 per cent of the total normative composition.

NORMATIVE COMPOSITION OF THE PERTHITES

The normative ratios Or:Ab and Ab:An of the feldspars shown in table 4

TABLE 2

ANALYSIS OF CLEAVELANDITE*

| Oxides | Per Cent Oxides |
|---|---|
| $SiO_2$ | 67.62 |
| $TiO_2$ | 0.00 |
| $Al_2O_3$ | 19.76 |
| $Fe_2O_3$ | 0.09 |
| MgO | 0.08 |
| CaO | 0.50 |
| BaO | 0.05 |
| $Na_2O$ | 11.03 |
| $K_2O$ | 0.09 |
| $H_2O$ | 0.40 |
| Total | 99.62 |

* Analyst, R. A. Higazy.

were plotted in figure 2. Perthite 1 has an Or:Ab ratio of 72.5:27.5, similar to that of perthite 13, which has the ratio 72.0:28.0. Yet the Ab:An is 88.5:11.5 in the former and 98.5:1.5 in the latter. The similarity of the Or:Ab ratio in the two perthites might indicate that their temperatures of formation were about the same, and, accordingly, one would expect them to have similar normative plagioclase composition. On the contrary, the normative plagioclase in the former is near oligoclase, whereas in the latter it is sodic albite. Similarly, perthite 14 and 6 have Or:Ab ratios of 73.1:26.9 and 73.2:26.8, respectively, and Ab:An ratios of 95.5:3.5 and 87.7:12.3, respectively. Moreover, figure 2 shows that there can exist perthites with similar normative plagioclase composition, but different Or:Ab ratios. Examples of this are: (1) perthites 15 and 8, which have Ab:An ratios of 97.0:3.0 and 96.5:3.5, respectively, and Or:Ab ratios of 70.7:29.3 and 81.3:18.7, respectively; (2) perthites 4, 5, and 6 with Ab:An ratios of 88.0:12.0, 87.4:12.6, and 87.7:12.3, respectively, and Or:Ab ratios of 82.0:18.0, 75.2:24.8, and 73.2:26.8, respectively.

Perthite 13, found as a phenocryst in the granites of the area, has a normative plagioclase composition of Ab 98.5. This is more sodic than most of the perthites. If this phenocryst had crystallized from a granitic magma, it would have a normative plagioclase poorer in the albite molecule than that of the pegmatite perthites because it formed at a comparatively higher temperature. Moreover, the normative plagioclase composition of cleavelandite (No. 18, table 4) is Ab 97.5 An 2.5. This is richer in anorthite than is the perthite phenocryst of the granite. Also, perthite 2 has exactly the same normative plagioclase composition as that of cleavelandite. Perthite 13 has even a lower anorthite content than does cleavelandite. Therefore, it can be stated that such perthites might have formed at the same temperature as cleavelandite (the

TABLE 3

CHEMICAL ANALYSES OF THE COUNTRY ROCKS*

| Oxides | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 68.83 | 70.85 | 69.93 |
| $TiO_2$ | 0.46 | 0.21 | 0.00 |
| $Al_2O_3$ | 14.63 | 15.48 | 17.84 |
| $Fe_2O_3$ | 1.42 | 0.64 | 0.14 |
| FeO | 3.83 | 1.09 | 0.20 |
| MnO | 0.05 | 0.01 | 0.01 |
| MgO | 2.24 | 0.83 | 0.54 |
| CaO | 0.52 | 0.38 | 0.59 |
| BaO | 0.11 | 0.11 | 0.02 |
| $Na_2O$ | 1.05 | 2.16 | 9.68 |
| $K_2O$ | 5.01 | 6.45 | 0.42 |
| $P_2O_5$ | 0.17 | 0.45 | 0.51 |
| $H_2O^+$ | 1.52 | 1.18 | 0.23 |
| $H_2O^-$ | 0.14 | 0.08 | 0.06 |
| Total | 99.98 | 99.90 | 100.17 |

* 1, Biotite schist; 2, microcline-perthite biotite schist; 3, albitite. Analyst, R. A. Higazy.

hydrothermal species), if not lower. Lastly, 17 is a microcline collected from the microcline-quartz zone of the Hugo pegmatite, and 16 is a perthite from the perthite zone. The perthite which is from an outer zone and supposed to have crystallized at an earlier stage than the microcline of the inner zone has a normative plagioclase composition of Ab 99.7 An 0.3; this is more sodic than that of the microcline, which has the normative plagioclase Ab 99.0 An 1.0.

### COMPOSITION OF PERTHITE OF THE SCHIST

The similarity in composition of the perthite metacryst in the schist and the phenocryst in the granite with the perthites of the pegmatites is striking. Table 5 shows that the weight percentages of all the oxides of a perthite meta-

TABLE 4

RATIO OR: AB: AN OF THE DIFFERENT PERTHITES*

| Number | Or | Ab | An | Ab:An | Per Cent Plag. | Or:Ab |
|---|---|---|---|---|---|---|
| 1 | 70.4 | 26.2 | 3.4 | 88.5:11.5 | 29.6 | 72.5:27.5 |
| 2 | 69.4 | 29.3 | 1.3 | 97.5: 2.5 | 30.6 | 70.2:29.8 |
| 3 | 94.7 | 3.5 | 1.8 | 66.0:34.0 | 5.3 | 96.2: 3.8 |
| 4 | 80.1 | 17.5 | 2.4 | 88.0:12.0 | 19.9 | 82.0:18.0 |
| 5 | 72.5 | 24.0 | 3.5 | 87.4:12.6 | 27.5 | 75.2:24.8 |
| 6 | 70.5 | 25.9 | 3.6 | 87.7:12.3 | 29.5 | 73.2:26.8 |
| 7 | 72.8 | 24.9 | 2.3 | 91.6: 8.4 | 27.2 | 74.5:25.5 |
| 8 | 80.7 | 18.6 | 0.7 | 96.5: 3.5 | 19.3 | 81.3:18.7 |
| 9 | 75.5 | 22.7 | 1.8 | 92.7: 7.3 | 24.5 | 76.8:23.2 |
| 10 | 85.3 | 13.3 | 1.4 | 90.6: 9.4 | 14.7 | 86.5:13.5 |
| 11 | 72.7 | 25.9 | 1.4 | 95.0: 5.0 | 27.3 | 73.7:26.3 |
| 12 | 79.0 | 19.3 | 1.7 | 92.0: 8.0 | 21.0 | 80.5:19.5 |
| 13 | 71.7 | 27.9 | 0.4 | 98.5: 1.5 | 28.3 | 72.0:28.0 |
| 14 | 72.1 | 26.6 | 1.3 | 95.5: 4.5 | 27.9 | 73.1:26.9 |
| 15 | 70.0 | 29.1 | 0.9 | 97.0: 3.0 | 30.0 | 70.7:29.3 |
| Average | 75.8 | 22.4 | 1.8 | 91.0: 9.0 | 24.2 | 77.2:22.8 |
| 16 | 71.1 | 28.8 | 0.1 | 99.7: 0.3 | 28.9 | 72.2:28.8 |
| 17 | 89.0 | 10.9 | 0.1 | 99.0: 1.0 | 11.0 | 89.0:11.0 |
| 18 | 0.6 | 97.0 | 2.4 | 97.5: 2.5 | 99.4 | 0.8:99.2 |

* 1–15, perthites; 16, perthite (Hugo pegmatite); 17, microcline (Hugo pegmatite); 18, cleavelandite.

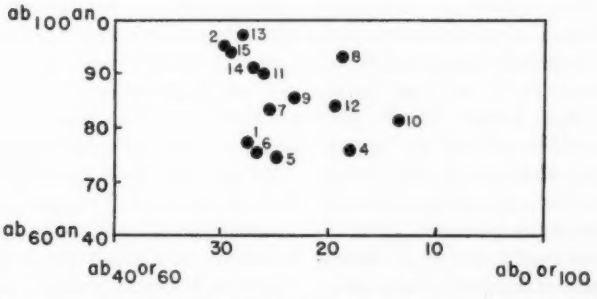


FIG. 2.—The normative ratios Or:Ab and Ab:An of the analyzed feldspars

cryst of the perthitized schist, with the single exception of baria and water, lie within the range of the oxides forming the perthites of the pegmatites. It also shows that the weight percentages of all the oxides of the perthite phenocryst of the granite, with the exception of lime, lie within the range of the oxides forming the perthites of the pegmatites.

TABLE 5

| Oxides | Range in the Perthites (Per Cent) | 1* | 2† |
|---|---|---|---|
| $SiO_2$ | 63.89–65.13 | 65.10 | 63.96 |
| $TiO_2$ | 0.00 | 0.00 | 0.00 |
| $Al_2O_3$ | 18.92–21.61 | 19.55 | 19.39 |
| $Fe_2O_3$ | 0.06– 0.13 | 0.10 | 0.13 |
| MgO | 0.00– 0.12 | 0.00 | 0.05 |
| CaO | 0.15– 0.82 | 0.07 | 0.46 |
| BaO | 0.01– 0.06 | 0.02 | 0.17 |
| $Na_2O$ | 0.38– 3.29 | 2.98 | 2.80 |
| $K_2O$ | 10.58–15.62 | 11.58 | 12.41 |
| $H_2O^+$ | 0.12– 0.31 | 0.16 | 0.36 |

* Perthite phenocryst in granite.
† Perthite metacryst in schist.

## TEXTURES OF PERTHITES

The analyzed perthites as well as a few others were examined microscropically. Four textural types have been recognized, namely,

1. Patch perthite
2. Vein perthite
3. Film perthite
4. String perthite

These types were named by Andersen (1928), who gave a complete description of each type. The interlocking perthite of Andersen (1928, p. 151), the braid perthite of Goldich and Kinser (1939), and the guttate perthite found by Barth (1930) were not observed in the Black Hills perthites. Two or more types of perthite always occur together. Table 6 shows the types found in the perthites analyzed. Those not analyzed chemically are similar to those analyzed.

Vein and patch types of perthite are abundant in practically all material studied. The patches reach 2.3 mm. or more in width and in most cases exceed 1.0 mm. In the patches of albite are islands of microcline, which exhibit complete optical continuity with the host microcline (pl. 2, *O*). They are well shown in sections parallel to both (010) (pl. 2, *E*) and (001) (pl. 2, *F*). Plates 2, *D–F*, and 3, *A* and *B*, illustrate these relations. In sections parallel to (001), regular bands of albite are commonly found which are parallel to the pericline twins of microcline and in many cases swell out perpendicular to this direction. Commonly the patches send out branches or offshoots which appear to have eaten their way into the microcline. These are irregular and in places have a complicated texture. In some cases the patches form lens-shaped or roughly rounded pools, which send off other patches which may run in any direction. The contacts between the albite of the patches and the microcline are commonly irregular and embayed (pl. 2, *E*). The amount of albite in the patches, estimated by microscopic areal measurements, is not the same in all perthites. In perthite No. 10 albite patches form about 8 per cent, whereas in perthite No. 11, albite patches form approximately 19 per cent of the whole specimen. Patch albite commonly shows polysynthetic twinning according to the albite law. The extinction angle on (010) varies, probably because of variation in the anorthite content. The microcline portions of the perthites are more homogeneous. The extinction angles of microcline are $+15°$ to $+16°$ on (001) and $+5°$ to $+6°$ on (010). The grating may differ in coarseness; in some cases it is fine, and in others it is relatively coarse. A combination of both fine and coarse gratings in a single perthite is common. Nearly all

the microcline seen contains some areas, variable in width, which lack the characteristic grating (pl. 2, *D*). Some graphic granites, composed essentially of quartz and perthite, also show a patch type which displays the general characteristics mentioned above.

The vein type of perthite is closely associated with the patch type. In some places it is difficult to differentiate between them. The vein perthite generally shows intricate and irregular networks of anastomosing veins, which grade into the patch type. The veins commonly enclose microcline islands in complete optical continuity with the host microcline similar to those in the patch type. Albite of the vein type rarely exhibits polysynthetic twinning, and its edges are commonly irregular. Both the width and the length of the albite of the vein type vary in the different perthites, but they are decidedly less than those of the patches. The width varies from 0.05 to 0.2 mm., and the length reaches a few millimeters. The veins in some cases are more or less parallel to (100) (pl. 3, *C*), but more commonly they run irregularly through the whole perthite. They generally have numerous veinlets which possess smaller width and length. Veins which have circular cross-sections of a diameter of about 0.05 mm. are also present (pl. 3, *D*).

TABLE 6

TEXTURAL TYPES OF ANALYZED PERTHITES

| No. | String | Film | Vein | Patch | Color* |
|---|---|---|---|---|---|
| 1 | T† | T | X | X | Pale pink |
| 2 | X | T | T | X | Grayish-white |
| 3 | T | T | X | X | Pink |
| 4 | ... | ... | X | X | Grayish-white |
| 5 | T | X | T | X | Deep pink |
| 6 | T | T | X | X | Pink |
| 7 | T | X | X | T | Salmon-red |
| 8 | ... | X | X | T | Whitish-gray |
| 9 | X | T | X | X | Grayish |
| 10 | ... | T | X | T | Grayish-white |
| 11 | T | T | X | X | Deep pink |
| 12 | T | ... | X | X | Whitish-gray |
| 13 | T | T | X | X | Pink |
| 14 | X | ... | X | T | Deep pink |
| 15 | ... | ... | X | X | Pinkish-gray |

* The perthites analyzed vary in color. The percentage of $Fe_2O_3$ may be the same in both pink and gray perthites. For example, perthites 1 and 2 are pink and grayish-white, respectively; yet each has 0.09 per cent $Fe_2O_3$. This indicates that the variation in color is not due to iron or apparently to any of the common elements but might be attributed to other rarer elements.

† Symbols: X = Present in abundance; T = Present in trace; ... = Absent.

The film and string albites constitute a relatively small amount of the total albite of the perthite, or both films and strings may be absent. The strings and films are always accompanied by vein and patch types (pl. 3, *A*). Both of the latter form, on the average, 90 per cent of the total albite of the perthite, the rest being film and string types. The total albite, estimated by microscopic areal measurements, conforms closely to the normative plagioclase percentages of the various analyzed perthites given in table 4. The films are generally 0.01 mm. in width, rarely 0.04 mm., and several millimeters long. The strings have a width

of about 0.003 mm. and a length of as much as 0.65 mm. Both film and string types commonly grade into the vein type. The strings make an angle of 76°–81° with (001). They commonly occur in groups between the veins. The grating of the microcline of the strings is relatively fine, but such fine grating can also be found in the other types.

### SIGNIFICANCE OF PERTHITE TEXTURES

The perthitic intergrowth has been thought to be due to (1) simultaneous crystallization, (2) exsolution, or (3) replacement.

A detailed study of the textural features of perthites of southern Norway led Andersen to believe that each well-defined textural class corresponds to a particular mode of origin (1928, p. 163). The string and film types were believed to be the products of exsolution. The string forms at an early stage of the evolution of the feldspar, whereas the films form at a lower temperature in a later stage. The vein perthite was believed to be formed mainly by recrystallization of albitic solutions derived from the pegmatitic magma from which the host feldspar crystallized. These solutions fill the contraction cracks, believed by Andersen to be formed perpendicular to (010) and making an angle of about 70° with (001), where they crystallized without much replacement. Goldich and Kinser (1939, p. 423) thought the atomic structure was an important factor in determining the location of the openings in which the vein solutions would crystallize. In a later stage, Andersen says, these solutions work their way farther from the cracks, replacing the host microcline and forming networks of albitic material. The patch type was considered to be an advanced stage of replacement, transitional between the vein type and pure albite.

It is reasonable to accept the replacement hypothesis for the explanation of the origin of the patch type. It is difficult to believe that magmatic crystallization can give rise to the peculiar characteristics of this type. The microcline islands in complete optical continuity with the host microcline, the extreme irregularity of the contact between the patches and the host, the numerous embayments, the network-like forms and other complex textures without any regularity in the direction of elongation, the transition of veins into patches and of the latter into chessboard albites and albite in general, the variation in the amounts of albite of the patches of different perthites, and the pools and pods of albite with offshoots and tongues which appear to have worked their way by eating and digesting the host—all these features undoubtedly favor the replacement theory.

Vein perthite, which is next to the patch type in abundance, commonly shows microcline relics and irregular and embayed contacts with the host microcline. Intricate networks of arborescent and anastomosing veins are common. These generally grade into the patch type. It is, therefore, a product of metasomatic processes, that is, a crystallization of albitic materials circulating into the contraction cracks and/or the openings controlled by the atomic structure of the feldspar. Much replacement of the potash feldspar accompanies its precipitation.

Andersen's concept of origin of string and film types of perthites seems correct. These perthites are the rarest types. They are believed formed by exsolution in the solid state, that is, by diffusion of the constituents of albite to cracks in the host microcline. Both types show more uniform elongation than do the vein and patch classes. Transitional stages be-

tween string and film types, on the one hand, and the vein type, on the other, are not uncommon. Some veins cut the strings and films and are therefore younger than such strings and films. In the early stages of the evolution of the perthite, diffusion of the albitic material seems to have followed a certain preferred direction in the atomic structure of the feldspar forming the regularity in trend of the strings and films. Later in the evolution some of the albitic materials acted selectively in certain directions in the feldspar, giving rise to the regular veins. They might have started this action in the same direction as those of the strings and films, transforming these into the vein type. Perthite No. 15 is devoid of both the string and the film types. This may be because the transformation of strings and films into veins has been completed.

What is the origin of the homogeneous potash-soda feldspar and of the circulating albitic material? It has been proved experimentally that soda feldspar dissolves to a considerable extent in potash feldspar by prolonged heating somewhat below the melting temperatures (Dittler and Köhler, 1925). It has also been proved by X-ray analyses that homogeneous potash-soda feldspar may form cryptoperthite on heating slightly below its melting temperature (Kozu and Endo, 1921; Chao, Smare, and Taylor, 1939). It may be concluded, therefore, that at relatively high magmatic temperatures the potash-soda feldspar is homogeneous. On cooling, the solubility of the soda feldspar decreases rapidly, so that a once homogeneous substance separates into two solid phases, namely, a soda-rich and a potash-rich feldspar. The soda feldspar in this case would be found in fine or thin intergrowths, in other words, strings and films, but not in relatively coarse bands, like those of the veins and patches. The perthite obtained experimentally is a submicroscopic cryptoperthite having a greater similarity to the strings, and to a less extent to the films, than to any of the other classes.

Although the string perthite can be produced by exsolution from magmatic sources, some may have been formed by migration of material from the country rock to structurally favored places where they form rocks markedly unlike the original parent-rock. This process is described elsewhere in this paper. The perthite of the perthitic schists (pl. 1, *A–D*) could not be derived from a magmatic source. It shows megascopic and microscopic features, discussed above, which are difficult to explain by viscous intrusion and magmatic crystallization.

## PERTHITE GENESIS

Magmatic pegmatites must, by definition, precipitate from the residual magmatic melt and are its final product of differentiation. The water content of this residual melt was probably less than 10–12 per cent. Such melts should obey the laws of crystal fractionation.

The concentration of lithium or sodium in certain pegmatites and the impoverishment of others in potassium seem to indicate that these rocks might not have been derived from the crystallization of a pegmatitic rest magma. The residual magma is known to be rich in the alkalies in the latest stages, and there is no way of separating the different alkali members by the differential crystallization of the liquor, letting one or another of them concentrate in certain more favorable structures.

In some places, as in the case of the Edison pegmatite in the Keystone area, potassium-bearing minerals are not found. This pegmatite has an interior

zone of quartz and albite, surrounded by a zone of quartz, albite, and spodumene. In other cases, such as the Etta pegmatite of the Keystone area, in which spodumene, the dominant mineral, forms crystals several feet in length, lithium is present in abundance; in still other pegmatites, lithium is found only in traces.

Whenever perthite forms pegmatitic pockets or forms a zone in a pegmatite body, its chemical composition should closely approximate the mother-liquor from which it was crystallized. Walker and Mathias (1946) believe that the perthite at Bantry Bay (South Africa) represents the composition of the pegmatitic magma. If perthites crystallize from a magma, such magma should follow the course of crystallization expressed by the feldspar equilibrium diagram.

The orthoclase-albite equilibrium diagram suggested by Vogt (1926) does not satisfactorily explain the derivation of the perthites analyzed from a granitic magma. The orthoclase:albite ratio in these perthites (table 4) is far from that which should be attained (that is, Or 40:Ab 60) had they been formed in such a manner.

Spencer (1938) in his study also concluded that the orthoclase-albite binary diagram cannot explain satisfactorily the magmatic origin of the potash-rich pegmatites.

### THE ORTHOCLASE-ALBITE-ANORTHITE TERNARY EQUILIBRIUM DIAGRAM

The ratio Or:Ab:An of the different perthites given in table 4 was plotted on the orthoclase-albite-anorthite equilibrium diagram as shown in figure 3. The cotectic curve, $CD_1$, in figure 3, after Bowen (1928), has $C$ at the Or:An eutectic point and $D$ at the Ab:Or eutectic point.

All the perthites analyzed lie in the field of the orthoclase or the potash-rich rocks. Those of Andersen (1928, p. 148) fall in the same field. In the normal course of crystallization, composition of the liquid cannot cross the cotectic curve. In the plagioclase field the crystallizing liquid must follow a certain path, such as that given by Nockolds (1946, p. 215) and plotted in figure 6 or that given by Barth (1944). The liquid, upon hitting the cotectic curve, would then follow it, so that in the last stages of crystallization the eutectic ratio Or:Ab would be attained. The Or:Ab:An ratio of the Black Hills rocks does not harmonize with this concept.

The author also plotted the Or:Ab:An content of some fifty rocks.[4] These are deep seated (graphic granites, granites, pegmatites, and a monzonite), hypabyssal (porphyries), and extrusive (rhyolites, liparites, pitchstones, comendites, and trachytes). The compositions of these fifty rocks also lie in the orthoclase field (fig. 4). An attempt to shift the Or:Ab eutectic point $D$ of figure 4 to $D^1$, around which most of the plotted graphic granites and granites fall, and the Or:An eutectic point $C$ to $C^1$, which represents the composition of a pitchstone,[5] would not solve the problem, as there are still other rocks[6] whose composition still lies in the orthoclase field (fig. 5). Why should these rocks, if of magmatic origin, not follow the regular course of crystallization that such an origin demands according to the feldspar ternary diagram?

[4] Of subrang Dopotassic Omeose, Order Quardofelic Britannare, and the subrang Dopotassic Magdeburgose and Sodipotassic Alaskose, Order Quarfelic Columbare, from Washington's tables (1917, pp. 109, 57, and 61, respectively).

[5] Belonging to the subrang Perpotassic, Order Quarfelic Columbare (Washington, 1917, p. 79).

[6] Of the subrang Dopotassic Omeose and the subrang Perpotassic Lebachose, Order Quardofelic Britannare (Washington, 1917, p. 107).

Bowen has proved (1937) that liquids at the latest stages of crystallization become enriched in the alkali silicates. The chemical analyses of some volcanic rocks have been shown to fall in the trough of rocks, Bowen referred to the work of R. Terzaghi (1935). She stated that some of the potash-rich rocks may have been formed by metasomatic alterations but that most of them are magmatic. The

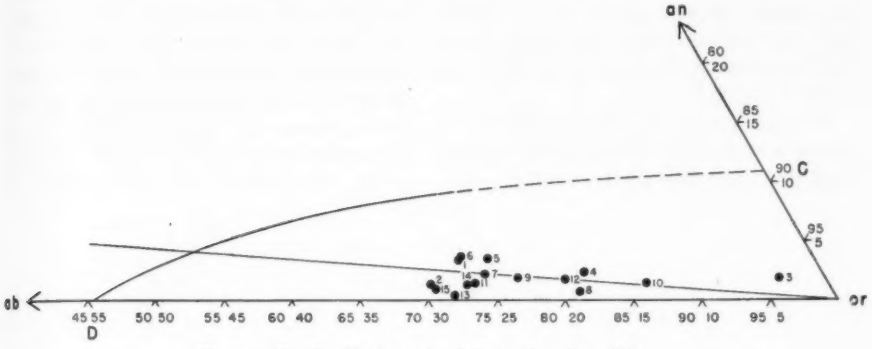


FIG. 3.—The Or:Ab:An ratio of the analyzed perthites

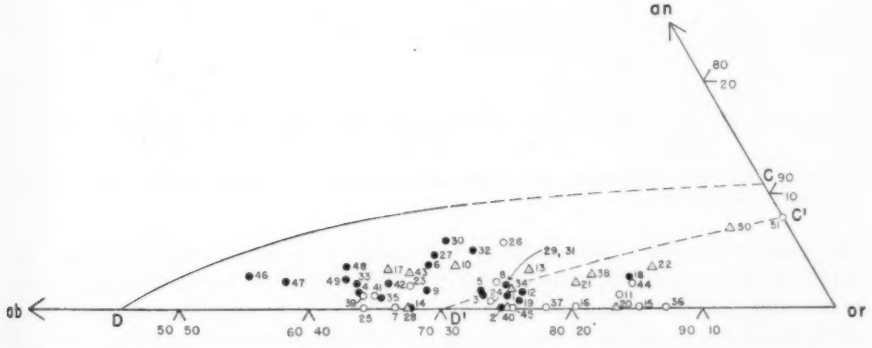


FIG. 4.—The Or:Ab:An ratio of some potash-rich rocks. Nos. 1–5 and 6–32 = nos. 1–5 and 7–33 of Washington (1917, p. 109). Nos. 33–45 = nos. 1, 2, 3, 11, 12, 13, 15, 18, 22, 25, 27, 30, and 31 of Washington (1917, p. 57). Nos. 46–49 = nos. 10, 31, 32, and 60 of Washington (1917, p. 61). Nos. 50–51 = nos. 1–2 of Washington (1917, p. 79). Solid circle = granite, pegmatite; open circle, rhyolite, trachyte, and other volcanics; triangle, porphyry.

his figure 9 (1937, p. 18), which he postulates for rocks formed in the last stages of crystallization. Some deep-seated rocks (granites and syenites) have been shown to lie also in this trough. However, most of the potash-rich rocks—extrusive as well as deep seated—do not lie in this trough. With respect to the origin of such position of the latter rocks in the orthoclase field Or:Ab:An ternary diagram she attributes mainly to the reaction relation between anorthite and orthoclase, to pressure, or to devitrification.

The reaction relation between anorthite and orthoclase as suggested by Bowen (1928, pp. 227–233) was thought

to enable the liquids to cross the cotectic curve. Recently, Schairer and Bowen (1947), in their study of the system anorthite-leucite-silica, found that such a reaction relation does not exist. There is, however, a eutectic between anorthite and orthoclase at 1040° ± 20° C. Therefore, the reaction factor can no longer be held responsible for liquids crossing the cotectic curve.

Terzaghi (1935) says that an increase of pressure under deep-seated conditions can shift the orthoclase-albite eutectic point; as rhyolites and liparites, as well as some deep-seated granites and pegmatites, lie together in the same field. Furthermore, Nockolds (1946) found that there are deep-seated aplites, adamellites, and granites which lie on the cotectic curve and follow its course (fig. 6). The Or:Ab:An ratios of some fifty deep-seated, hypabyssal, and extrusive rocks,[7] plotted in figure 6, lie approximately on the cotectic curve, and some of them fall in close proximity to the Or:Ab eutectic point. Such types of rocks might be

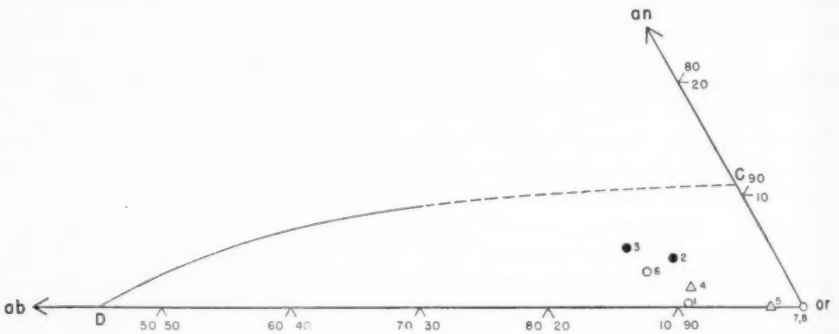


FIG. 5.—The Or:Ab:An ratio of some potash-rich rocks. Nos. 1–8 = nos. 4, 5, 7, 8, 9, 10, 11, and 12 of Washington (1917, p. 107).

she admits, however, that, in order to establish this, more information is needed concerning the latent heat of formation of the feldspars. If such were the case, the plagioclase field would increase, and some of the potash-rich rocks would no longer lie in the orthoclase field. But, as has been mentioned above (p. 566), such a shift, on the basis of the position of most of the graphic granites (which closely approximates the amount of shift postulated by Terzaghi), is not adequate because other potash-rich rocks will still remain in the orthoclase field. The effect of pressure on the feldspar equilibrium diagram appears to be of little importance because some volcanic rocks, such mainly of magmatic origin. Furthermore, there should be pegmatites corresponding to such rocks with compositions approximately on the cotectic curve, so that the position of this curve as given by Bowen in the orthoclase-albite-anorthite equilibrium diagram seems to be approximately correct.

Therefore, if pressure was not the factor responsible for throwing potash-rich rocks into the orthoclase field, would it not also have affected Nockolds' selected rocks and the others of figure 6 which do not lie in that field? Why, then,

[7] Of the subrangs Sodipotassic Liparose, Dopotassic Mihalose, and Dopotassic Dellenose (Washington, 1917, pp. 145, 79, and 165, respectively).

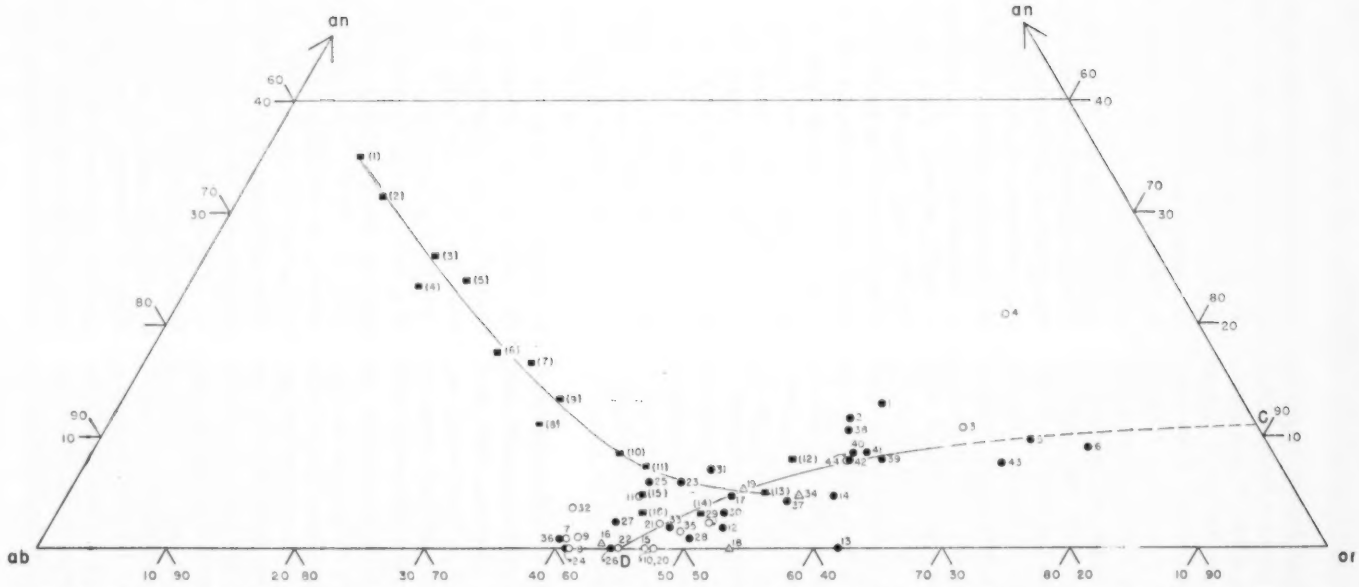


FIG. 6.—The Or:Ab:An ratio of some rocks following a course of crystallization in the plagioclase field and of rocks falling approximately on the cotectic curve. Nos. (1)–(16) = nos. 1–16 of Nockolds (1946). Nos. 1–6 = nos. 3–8 of Washington (1917, p. 79). Nos. 7–37 = nos. 248, 249, 250, 254, 255, 256, 257, 258, 260, 261, 262, 263, 264, 265, 267, 270, 276, 279, 280, 282, 283, 284, 285, 289, 290, 291, 292, 294, 296, and 297 of Washington (1917, p. 145). Nos. 38–44 = nos. 1, 5, 7, 8, 10, 17, and 22 of Washington (1917, p. 165). Square, granite, adamellite, and aplite of Nockolds (1946).

are they still governed by the magmatic rules? Why should pressure have affected some rocks and not others? If one argues that rocks studied by Nockolds, though deep seated, were subjected to less pressure than were the deep-seated rocks in the orthoclase field, then it follows that the volcanic rocks in the orthoclase field also were formed under conditions of higher pressure than the rocks of Nockolds—a conclusion obviously absurd. There probably are deep-seated rocks—granites and aplites, as well as pegmatites—which are unquestionably of magmatic origin; but these will generally have normative feldspar contents which follow the cotectic curve, as in the case of those rocks selected by Nockolds. At the same time, there are other deep-seated varieties of the same species which do not follow the magmatic crystallization rule. The orthoclase-albite-anorthite ratios of such rocks cannot result from strictly magmatic processes; thus metasomatic alterations must have played the significant role in their formation.

It is important to note that the zoned plagioclase crystals of an extrusive rock formed under low pressure, such as a basalt, and those of a deep-seated rock formed under relatively high pressure, such as a gabbro, show agreement with the course of crystallization given in the the albite-anorthite solid-solution diagram (Bowen, 1928, p. 34). In both rocks the zoned crystals have cores which are richer in the anorthite molecule than are the rims. Moreover, the path of crystallization of both basalts and gabbros terminates at the same point in the diopside-albite-anorthite equilibrium diagram (Bowen, 1928, p. 46). Diopside and albite exhibit quite different crystal structures: chain lattice and framework lattice, respectively. It seems probable that pressure would have a greater effect on systems whose members have different crystal structures. Because no pressure effect is evident in the albite-anorthite and the diopside-plagioclase systems, it must be negligible in the orthoclase-albite-anorthite system.

Terzaghi (1935) found that four devitrified rhyolites were richer in potash and poorer in both soda and lime than were rhyolites not devitrified. The Or: Ab: An ratios of the devitrified rocks fall in the potash-rich or orthoclase field. Terzaghi believed that the devitrified rhyolites were originally magmatic and vitric and that at an earlier stage of their history they had the normal Or: Ab: An ratios of vitreous rhyolites. Later, soda, lime, and potash were leached out—the soda and lime at a greater rate than the potash. Therefore, when these altered rocks became devitrified, they possessed higher potash content than did the original vitric type, and they consequently fell into the orthoclase field. Recently the plot of the normative feldspar content of some rhyolite rocks of the Esterel region in France (Terzaghi, 1948, fig. 1) shows that the vitreous obsidian lies in the plagioclase field, whereas six devitrified obsidians and spherulitic rhyolites and two rhyolite porphyries lie in the orthoclase field. Because devitrification is confined to the immediate vicinity of the cracks in one of the obsidians, Terzaghi believes the devitrification to be due to percolating solutions. The crystals of alkalic feldspar, which are most strongly euhedral, are believed to have originated by *replacement* of vitreous material. Terzaghi concludes that hot solutions and possibly gaseous emanations were active in the region after the eruptions of rhyolitic lava took place. Reactions between solutions and rocks are most clearly indicated by the microscopic features

of the spherulitic rhyolites that she studied. The secondary potash which enriched the rhyolites during their devitrification either was introduced by hydrothermal solutions or gaseous emanations originating from an underlying body of magma or was derived by local leaching of the rhyolite itself. The potash leached from the rhyolite in one place was carried on in solution and was available for potash enrichment elsewhere. The possible dual origin of the secondary potash in the Esterel region is in harmony with the conclusions reached by Allen (1935), Allen and Day (1934), and Fenner (1936) in their studies of the thermal waters and the altered rhyolites of the Yellowstone Park region.

Fenner (1936, p. 278) observed that not only were orthoclase, analcite, heulandite, and calcite formed as replacements of original minerals by soluble salts (chlorides, carbonates, etc., of the alkalies), but all these difficultly soluble minerals were taken into solution to some extent, migrated rather freely, and were deposited in fissures and other open spaces. Elsewhere, Fenner (1936, p. 279) states that there may be many occurrences of similarly altered rocks in volcanic regions, which, in the absence of complete data, have been regarded as normal lavas. Even analyses of such rocks have been accepted as representing original compositions.

The high potash content of some rhyolites and extrusives may be due to secondary enrichment by potash-bearing solutions. The secondary potash might have been derived in part from these extrusives by leaching and migrating from one place to another. Possibly the potash transfer could take place in the solid state, especially so if evidence for the presence of solutions is lacking. If pressure causes differences in the chemical potentials of the elements in the various parts of country rocks, it is believed that the elements will migrate from areas of higher to areas of lower potentials until equilibrium is reattained (Ramberg, 1944).

Other factors may be responsible for potash-rich pegmatites—perhaps the concentration of volatiles in the last stages of crystallization. Andersen (1931, p. 25) gives two analyses of pegmatites whose projection points apparently lie in the plagioclase field. Others can be found in Washington's tables. On the other hand, some pegmatites lie in the orthoclase field. Though all magmatic pegmatites are supposed to have crystallized from residual solutions rich in volatiles, we find some on the cotectic curve, others in the plagioclase field, and still others in the orthoclase field. We can conclude, therefore, only that we have pegmatites of different origins and that in the case of the magmatic types the presence of volatile constituents has had no significant influence on the course of crystallization.

Moreover, the effect of volatile constituents on the course of crystallization is not too well known; their only important effect, however, seems to be a lowering of the temperature of crystallization, with no discernible influence on its course. Bowen states (1928, pp. 299–300) that the addition of 1 per cent $H_2O$ (the most important volatile constituent) to a liquid composed of 50 per cent $K_2Si_2O_5$ and 50 per cent $K_2SiO_3$ does not change the course of crystallization or even the proportions of the separated phases from that of the dry melt. Moreover, in Bowen's nephelite-kaliophyllite-silica equilibrium diagram (1937), the lowest temperature of crystallization for a dry melt is 1038° C.; but granites rich in the volatiles and thought to have crys-

tallized from a magma at 700° C. have chemical compositions which place them in the "trough" of this dry-melt diagram. The zoned plagioclase crystals of both peridotites poor in volatiles and granites relatively rich in these follow equally well the rules of plagioclase crystallization given by the albite-anorthite binary equilibrium diagram. One is therefore led to the conclusion that, although the presence of the volatile constituents will lower crystallization temperatures, it will not significantly change the course of crystallization. That is, although fusion surfaces might show differences in relief with increase or decrease of volatiles, the positions and relative reliefs of the cotectic "valleys" and boundary surfaces will remain approximately the same.

The immiscibility of water-rich siliceous liquids does not seem to be the solution. The initial compositions used by Tuttle and Friedman (1948) for the determination of the equilibrium relations of the system $Na_2O$-$SiO_2$-$H_2O$ at 250, 300, and 350°, proving that the system possesses liquid immiscibility between 200 and 250°, consist of approximately 50 per cent $H_2O$. This is much higher than the 10 per cent of $H_2O$ assumed to be in the pegmatitic rest magma. Smith (1948, p. 538) also dealt with solutions of a very highly specialized composition that do not represent a rest pegmatitic magma from which a perthitic rock could crystallize. Smith's solutions had 50 per cent $H_2O$ and less than 1 per cent of $K_2O$. Goranson's work (1931) makes it unlikely that a perthitic liquid which approximates that of a granite could absorb more than 10 per cent $H_2O$.

Moreover, the potash-rich rhyolites cannot be explained because their temperature of crystallization is higher than that at which immiscibility occurs.

The experiments by Goranson (1931) have demonstrated a limited miscibility of $H_2O$ in silicate melts. A consequence of this is that, near the end of the consolidation history of a magma, a hydrothermal solution may separate as an independent phase coexisting with the magmatic phase, on the one hand, and with the solid phase (the rock), on the other. The significance of Goranson's work and its implications have been discussed by many petrologists. The exudation of the hydrothermal solutions is believed to mark the final stages in the consolidation of the magma. It is important to note that they are separated from the magma through an immiscibility gap. The composition of the solutions is governed by the endomagmatic hydrothermal differentiation processes as described by Neumann (1948). Magmatic pegmatites are the precipitates from the residual magmatic melt prior to the hydrothermal stage.

It appears, therefore, that neither the reaction relation anorthite-orthoclase nor pressure nor the volatile constituents can account for the composition of the Black Hills perthitic pegmatites. Metasomatic alterations which were found by Terzaghi (1948) to be responsible for the presence in the orthoclase field of the potash-rich rhyolites that she studied might also be the reason why the Black Hills potash-rich pegmatites lie in the same field. It is uncertain whether all the potash-rich deep-seated, hypabyssal, and extrusive types plotted (figs. 4 and 5) are metasomatic rocks or were formed by some other processes. This is not discussed in this paper and will be the subject matter of another work by the author.

If it turns out that the hypothesis of potassium metasomatism is true, the orthoclase field of the orthoclase-albite-

anorthite ternary diagram might well be named the "potassium metasomatism field." Perhaps also there are rocks which possess a higher albite normative composition than is consistent with a truly magmatic origin. Examples of these are the albite-rich granites and the albitites. A chemical analysis of an albitite from the Black Hills is given in table 3; this has an Or:Ab ratio of 3:97. These would be placed in a field nearer to the Ab corner of the Or-Ab-An ternary diagram; this field might be named the "sodium metasomatism field."

Whether the potash pegmatites of the Black Hills were once magmatic and have been metasomatized or are merely metasomatic is another subject and will be considered elsewhere.

### GENESIS OF PERTHITE SCHISTS

Table 3 shows the chemical analyses of three country rocks. The rock formula introduced by Barth (1948) of each of the analyzed rocks—the average perthite, the cleavelandite, and the microcline—was calculated as shown in table 7. Table 8 shows quantitatively the gains and losses required to change the normal schist to a perthitized schist, albitite, and perthite. The normative compositions, using Von Wolff's method, were also calculated in table 9.

Figure 7 is a reproduction of figures 7 and 8 of Reynolds' work on the geochemical changes leading to granitization (1946). The schist of the Glendale area lies in the field of the aureole sediments of the Flamouth granite area. It lies also in the field of the aureole shales of the Dartmoor area (Reynolds, 1946, fig. 5). The perthitized schist falls in the fields representing the Flamouth granite, the granitization series of the Dartmoor area, and the granites of the latter region. The albitite and the average perthite lie in a separate field closer to the *L* of the diagram. This means that partial feldspathization of the schist could give rise to perthitic schist, which has a chemical composition similar to potash granites. If feldspathization is more complete, rocks will form that possess a chemical composition near that represented by *L* of the Von Wolff diagram.

Such rocks are the albitites and the perthitites (monomineralic rocks composed of perthite). The formation of such rocks is an end-stage in Reynolds' sequence of changes of pelitic and semipelitic rocks. These changes may take place by the reorganization of the atoms and molecules inside the schist. Table 8 would then refer to loss and gain inside certain volumes of the schist, whereas the whole schist complex may show little change in bulk chemical composition. The perthite crystals in this schist seem to grow and recrystallize at the expense of the schist material, similar to the formation of porphyroblasts in other metamorphic rocks. In places they seem to have pushed aside the schist which forms the groundmass of the rock (pl. 1, *A*). In places there are crystals exceeding 6 cm. in length, exhibiting megascopic sieve texture (pl. 1, *C*). A perthitic pegmatite which has approximately the same chemical composition and shows identical textures as those of the perthite of the schist may thus be formed in a similar manner. It is worth mentioning that Goldschmidt (1934) attributed the source of scandium, the rare element found in some pegmatitic pockets, to the country rocks which are mainly amphibolites and not to a pegmatitic magma. This was concluded from spectrographic studies. The amphibolites which could give their scandium lend support to the belief that the schist could contribute potassium to form the feldspars. If this

be true, the change is illustrated by the following: Three schist "unit cells" (the least number of the schist "unit cell" to form a perthite is 3) are composed of: K 16.86, Na 5.37, Ca 1.62, Mg 8.76, Fe 11.43, Al 45.45, Ti 0.90, Si 181.35, P 0.39, H 26.85; and the average perthite "unit cell" consists of: K 14.45, Na 4.35, Ca 0.39, Mg. 0.05, Fe 0.07, Al 21.20, Ti 0.0, Si 58.80, H 1.37.[8] Therefore, the equation of the process of transformation will be

$$3 \text{ schist} \rightleftharpoons \text{perthite} + 2 \text{ schist relic.}^{9} \quad (1)$$

The "unit cell" of this schist relic is composed of: K 1.21, Na 0.51, Ca 0.62, Mg 4.36, Fe 5.68, Al 12.13, Ti 0.45, Si 61.28, P 0.20, H 12.74 (see Table 10).

Figure 8 shows a crystal of perthite about 2 cm. in length, forming a zone surrounded by another zone containing

[8] Ba is taken into Ca, and Mn is taken into Fe.

[9] Material produced after the petroblastic formation of perthite from the schist.

TABLE 7

FORMULAE OF THE COUNTRY ROCKS

| Rock | K | Na | Ca | Mg | Ba | Mn | $Fe^{++}$ | $Fe^{+++}$ | Al | Ti | Si | P | O | H |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Schist | 5.62 | 1.79 | 0.50 | 2.92 | 0.04 | 0.04 | 2.82 | 0.95 | 15.15 | 0.30 | 60.45 | 0.13 | 160 | 8.95 |
| Perthitic schist | 7.20 | 3.80 | 0.38 | 1.46 | 0.04 | 0.00 | 0.78 | 0.42 | 15.70 | 0.16 | 61.62 | 0.32 | 160 | 6.85 |
| Albitite | 0.48 | 16.30 | 0.54 | 0.70 | 0.00 | 0.00 | 0.15 | 0.10 | 18.34 | 0.00 | 60.51 | 0.38 | 160 | 1.32 |
| Average perthite | 14.45 | 4.35 | 0.37 | 0.05 | 0.02 | | | 0.07 | 21.20 | 0.00 | 58.80 | | 160 | 1.37 |
| Cleavelandite | 0.11 | 18.37 | 0.47 | 0.11 | 0.01 | | | 0.06 | 20.24 | 0.00 | 59.21 | | 160 | 2.42 |
| Microcline | 17.50 | 2.15 | 0.02 | 0.03 | | | | 0.03 | 20.90 | 0.00 | 59.00 | | 160 | 1.65 |

TABLE 8

GAINS AND LOSSES IN THE SCHIST

| | Addition and Subtraction |
|---|---|
| Schist to perthitic schist | *Addition:* 1.58 K, 2.01 Na, 0.55 Al, 1.17 Si, 0.19 P |
| | *Subtraction:* 0.12 Ca, 1.46 Mg, 0.04 Mn, 2.57 Fe, 0.14 Ti, 2.10 H |
| Schist to albitite | *Addition:* 14.51 Na, 0.04 Ca, 3.19 Al, 0.06 Si, 0.25 P |
| | *Subtraction:* 5.14 K, 2.22 Mg, 0.04 Ba, 0.04 Mn, 3.52 Fe, 0.30 Ti, 7.63 H |
| Schist to average perthite | *Addition:* 8.83 K, 2.56 Na, 6.05 Al |
| | *Subtraction:* 0.13 Ca, 2.87 Mg, 0.02 Ba, 0.04 Mn, 3.70 Fe, 0.30 Ti, 1.65 Si, 0.13 P, 7.58 H |

TABLE 9

NORMATIVE COMPOSITION AFTER VON WOLFF

| Rock | Alk $f$ | An | Mt | Aug | $Q$ | $t$ | $L$ | $M$ | $Q$ |
|---|---|---|---|---|---|---|---|---|---|
| Schist | 37.6 | 2.4 | 1.2 | 13.4 | 41.1 | 4.3 | 44.3 | 14.6 | 41.1 |
| Perthitic schist | 55.2 | 2.0 | 0.6 | 5.2 | 34.3 | 2.8 | 60.0 | 5.8 | 34.3 |
| Albitite | 84.2 | 2.8 | 0.1 | 2.0 | 10.9 | 0.2 | 87.2 | 2.1 | 10.9 |
| Average perthite | 94.7 | 3.9 | 0.1 | 0.1 | 0.1 | 1.1 | 99.7 | 0.2 | 0.1 |

schist relic material. Figure 9 is drawn to explain a mode of formation of such a crystal. The arrows indicate the direction of migration, and the amounts of

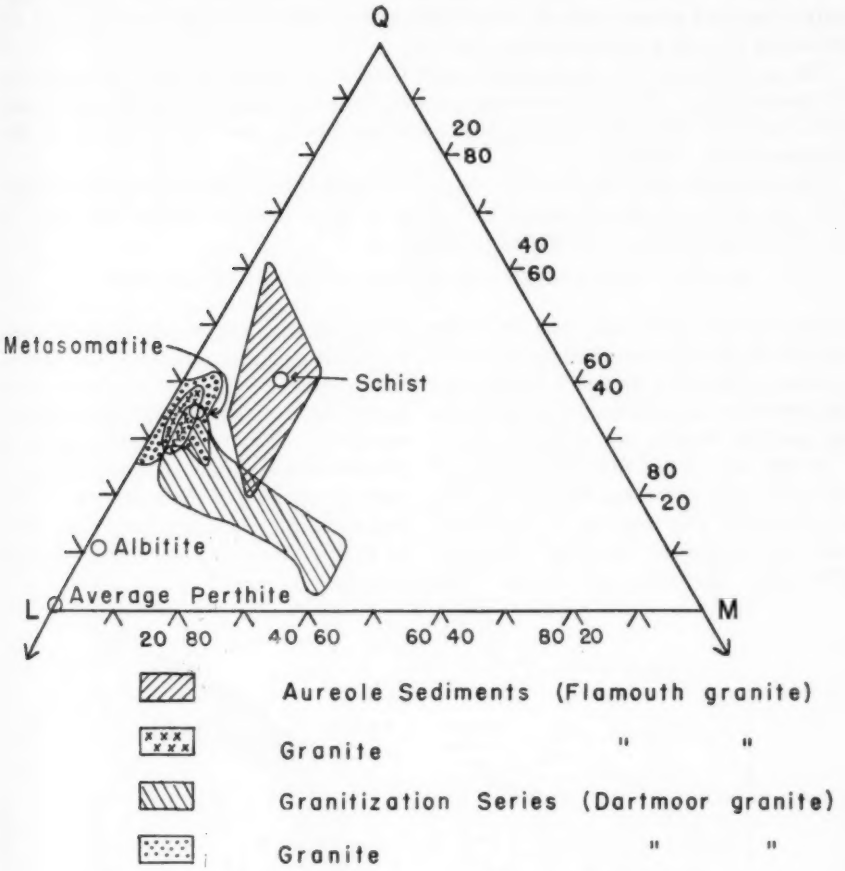


FIG. 7.—Von Wolff diagram showing the composition of the studied rocks

the diffusing ions are shown. The three "unit cells" of equation (1) are called in figure 9 the "outer shell," the "inner shell," and the "core." Potassium and sodium migrated from both the outer and the inner shells toward the core, which acted as a receiver for the feldspathic material. The core, which possesses K 5.62 and Na 1.79, has thus gained K 5.62 and Na 1.79 from the inner shell and K 3.21 and Na 0.77 from the outer shell. The core also gained Al 6.05 migrating from the inner shell and at the same time lost the excess of other ions over and above those necessary to form the perthite. The inner shell has gained K 1.21, Na 0.51, and Al 3.03 migrating from the outer shell and the other ions

migrating from the core in the amounts indicated on the arrows. Lastly, the outer shell gained material migrating from the core as given in figure 9. In this case it is supposed that the outer and the inner shells have the same composition after the development of the perthite. If such were not the case, the outer and inner shells would form two zones different in composition. The development of a pegmatite body composed almost entirely of perthite surrounded by schist material may be the result of such a transformation process.

The perthite zone of a pegmatite might form in a similar manner. But equation (1) does not apply if several zones are present. If the ratio and chemical composition of the different minerals forming these zones are known, one can set up a similar transformation equation.

The composition of cleavelandite (table 2) is considered to represent that of albite. It is possible for both microcline and albite to form from the schist according to the following equation (illustrated in table 11):

$$16 \text{ schist} \rightleftharpoons 4 \text{ microcline}^{10} + \text{albite (cleavelandite)} + 11 \text{ schist relic} . \qquad (2)$$

Sixteen schist "unit cells" are the least number of these units which could form a potash-rich and a soda-rich feldspar in the approximate ratio found in the average perthite, that is, 4:1, respectively.

In this case, if conditions in the country rock were favorable for zoning, the schist could give rise to a pegmatite with two zones, one composed of microcline and the other of albite. This transformation might take place at lower temperatures than those of the formation of perthite. This is because, at relatively high temperatures, the feldspathic material is likely to form a homogeneous potash-soda feldspar. It should be noted that the formula of the schist relic of the last equation is approximately the same as that of the schist relic derived from equation (1).

TABLE 10

| | 3 Schist = Perthite | + 2 Schist Relic | Formula of Schist Relic |
|---|---|---|---|
| K | 16.86 | 14.45 | 2.41 | 1.21 |
| Na | 5.37 | 4.35 | 1.02 | 0.51 |
| Ca | 1.50 | 0.39 | 1.13 | 0.57 |
| Mg | 8.76 | 0.05 | 8.71 | 4.36 |
| Ba | 0.12 | 0.02 | 0.10 | 0.05 |
| Mn | 0.12 | ..... | 0.12 | 0.06 |
| $Fe^{++}$ | 8.46 | ..... | 8.46 | 4.23 |
| $Fe^{+++}$ | 2.85 | 0.07 | 2.78 | 1.39 |
| Al | 45.45 | 21.20 | 24.25 | 12.13 |
| Ti | 0.90 | 0.00 | 0.90 | 0.45 |
| Si | 181.35 | 58.80 | 122.55 | 61.28 |
| P | 0.39 | ..... | 0.39 | 0.20 |
| H | 26.85 | 1.37 | 25.48 | 12.74 |

[10] The formula of microcline (table 1) of Hugo pegmatite is used here.

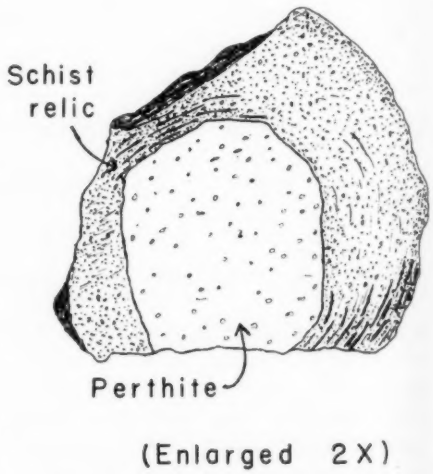


FIG. 8.—Perthite metacryst in the schist

FIG. 9.-
B, inner sh

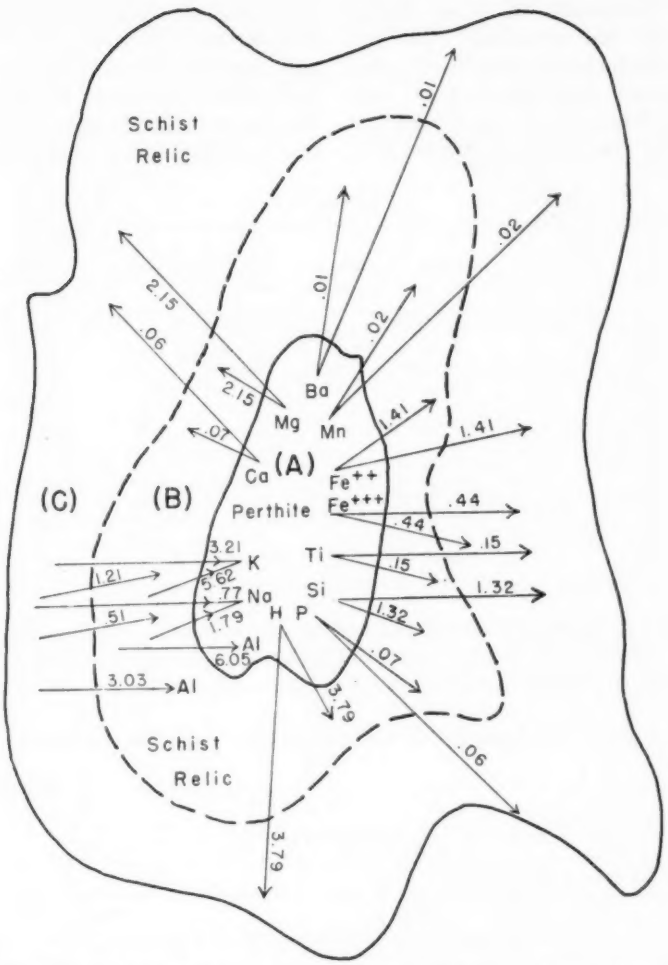


FIG. 9.—Reorganization of the schist material to form the perthite metacryst. *A*, core of innermost shell; *B*, inner shell; *C*, outer shell.

It is worth mentioning that the formation of microcline and albite suggests an explanation of the plagioclase rims surrounding orthoclase in some potash granites and the orthoclase rims surrounding plagioclase in some basic rocks. Such rimmed plagioclases were considered by Bowen (1928, pp. 227–233) as evidence for the existence of a reaction relation between anorthite and orthoclase. In a recent work, Schairer and Bowen (1947) found that anorthite and orthoclase are not reaction pairs; therefore, Bowen's first conclusion is no longer valid, and the rimmed plagioclases of the basic rocks must now be explained in another way. They could find their solution in solid diffusion phenomena.

TABLE 11

| | 16 Schist | = 4 Microcline | + Cleavelandite | + 11 Schist Relic | Formula of Schist Relic |
|---|---|---|---|---|---|
| K | 89.92 | 70.00 | 0.11 | 19.81 | 1.80 |
| Na | 28.64 | 8.60 | 18.37 | 1.67 | 0.15 |
| Ca | 8.00 | 0.04 | 0.47 | 7.49 | 0.67 |
| Mg | 46.72 | 0.12 | 0.11 | 46.49 | 4.22 |
| Ba | 0.64 | ..... | ..... | 0.64 | 0.06 |
| Mn | 0.64 | ..... | ..... | 0.64 | 0.06 |
| $Fe^{++}$ | 45.12 | ..... | ..... | 45.12 | 4.10 |
| $Fe^{+++}$ | 15.20 | 0.12 | 0.06 | 15.02 | 1.40 |
| Al | 242.40 | 83.60 | 20.24 | 138.60 | 12.60 |
| Ti | 4.80 | 0.00 | 0.00 | 4.80 | 0.43 |
| Si | 967.20 | 236.00 | 59.21 | 671.99 | 61.09 |
| P | 2.08 | ..... | ..... | 2.08 | 0.19 |
| H | 143.20 | 6.60 | 2.42 | 134.20 | 12.18 |

If microcline (potash-rich material) and albite (soda-rich material) derived from the schist subsequently reacted metasomatically by replacing the schist, other rocks would result, according to the following equation (illustrated in table 12):

Microcline + albite (cleavelandite) + schist $\rightleftharpoons$ perthite + albitite + perthite schist + basic ions . (3)

TABLE 12

| | Microcline | + Cleavelandite | + Schist | = Perthite | + Albitite | + Perthite Schist | + Basic Ions |
|---|---|---|---|---|---|---|---|
| K | 17.50 | 0.11 | 5.62 | 14.45 | 0.48 | 7.20 | ..... |
| Na | 2.15 | 18.73 | 1.79 | 4.35 | 16.30 | 3.80 | ..... |
| Ca | 0.02 | 0.47 | 0.50 | 0.37 | 0.54 | 0.38 | ..... |
| Mg | 0.03 | 0.11 | 2.92 | 0.05 | 0.70 | 1.46 | 0.85 |
| Ba | ..... | 0.01 | 0.04 | 0.02 | 0.00 | 0.04 | ..... |
| Mn | ..... | ..... | 0.04 | ..... | 0.00 | 0.00 | ..... |
| $Fe^{++}$ | ..... | ..... | 2.82 | ..... | 0.15 | 0.78 | 1.89 |
| $Fe^{+++}$ | 0.03 | 0.06 | 0.95 | 0.07 | 0.10 | 0.42 | 0.45 |
| Al | 20.90 | 20.24 | 15.15 | 21.20 | 18.34 | 15.70 | ..... |
| Ti | 0.00 | 0.00 | 0.30 | 0.00 | 0.00 | 0.16 | 0.14 |
| Si | 59.00 | 59.21 | 60.45 | 58.80 | 60.51 | 61.62 | ..... |
| H | 1.65 | 2.42 | 8.95 | 1.37 | 1.32 | 6.85 | ..... |

The amounts of potassium, sodium, aluminum, and silica of the reactants are approximately equal to those of the products; however, basic ions are released in the reaction and are thus free to migrate to other parts of the country rock, where they may, by reaction, give rise to other basic rocks. Perhaps they migrated to the north of the studied area, where they may have played a role in the formation of some amphibolites outcropping there. Another working hypothesis is that they might have diffused to lower levels in the gravitational field. A combination of both explanations is also a possibility. Figure 10 is a summary of the reactions discussed above.

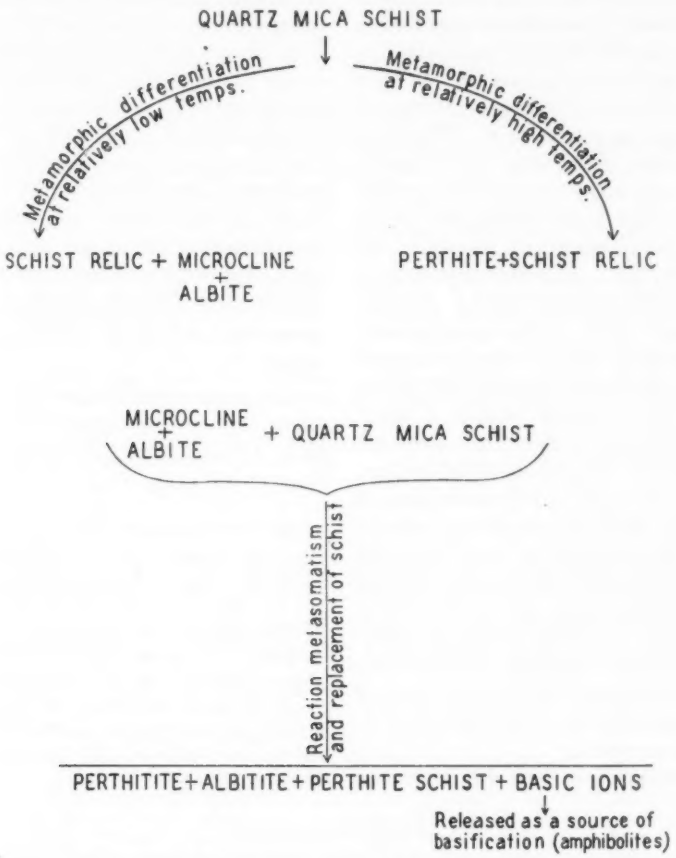
QUARTZ MICA SCHIST
Metamorphic differentiation at relatively low temps.
Metamorphic differentiation at relatively high temps.
SCHIST RELIC + MICROCLINE + ALBITE
PERTHITE+SCHIST RELIC
MICROCLINE + ALBITE
+ QUARTZ MICA SCHIST
Reaction metasomatism and replacement of schist
PERTHITITE+ALBITITE+PERTHITE SCHIST + BASIC IONS
Released as a source of basification (amphibolites)

FIG. 10.—Metasomatic development of the Black Hills perthite

## SUMMARY AND CONCLUSIONS

Some pegmatitic perthites and country rocks of the Glendale area of the Black Hills of South Dakota were chemically analyzed and microscopically studied in order to explain the origin of the perthites of this area. The textures of

the perthites and the microscopic features of the country rocks do not seem to be of a magmatic nature; rather, they support metasomatic activities. Evidence of metasomatic derivation of the perthites was given in the discussion of their chemistry. The feldspar equilibrium diagram was proved to be unsatisfactory for the magmatic derivation of these perthites. Moreover, the general chemistry does not harmonize with the general crystallization rules. The schists grade into feldspar-rich types, the country schists having been chemically reactivated. The feldspathic materials are thought to be endogenetically derived from the schist. Perthitization was partial in some parts of the country rock, where perthite schists were developed. In other parts of the country rock perthitization was complete because of favorable conditions of temperature, pressure, and structure, and thus perthitic pegmatite bodies formed. In still other parts of the country rock the conditions assisted both in the trapping of some elements and in the action of certain reaction processes which gave rise to the formation of shells or zones composed essentially of definite minerals characteristic of such conditions. The perthite zone of certain pegmatities is considered to be one of these zones.

The formation of perthite in general was a complex metasomatic process which took place in various stages. The following steps are believed to have taken place during the evolution of the studied perthites: (1) endogenetic development of homogeneous potash-soda feldspathic material by metamorphic differentiation of the country rocks; (2) diffusion of the soda feldspar portion into preferred parts of the structure of the homogeneous material to form the string and film types; (3) endogenetic development of a soda-feldspar (albite) from the schist; and (4) reaction metasomatism or replacement of microcline of the film and string types of perthite, giving rise to vein and patch classes. This reaction was accelerated by water derived from the schist.

I do not pretend to have discussed the whole problem of the genesis of pegmatites. I have given only some explanations of the studied types.

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# SANDSTONE TYPES: THEIR ABUNDANCE AND CEMENTING AGENTS[1]

SEFTON L. TALLMAN
University of Chicago

ABSTRACT

A selected sample of 277 sandstones (183 from thin sections and 94 from descriptions in the literature) were classified into five general classes on the basis of detrital mineral composition. The relative abundance of each of these classes was determined, as well as the average mineral composition of each class. A comparison of the carbonate-cemented sandstones with the silica-cemented varieties showed an increase in carbonate in later geologic periods. In the carbonate-cemented sandstones, partial replacement of the detrital grains was almost universal.

## INTRODUCTION

This study had two objectives. First, an attempt was made to measure the relative abundance of each of the major types of sandstones. Second, the variation with respect to time of the ratio of the silica-cemented sandstones to those carbonate-cemented, was studied.

Thin sections of the specimens were examined microscopically to determine specifically the detrital mineral components and the cementing agent and, incidentally, to record some of the structure of the rock, including such phenomena as replacement.

## ACCUMULATION OF THE DATA AND METHOD OF TABULATION

The initial problem of this study was to obtain a sample which would be representative of all sandstones. In an effort to accomplish this, the author secured 277 specimens of wide geographic and stratigraphic range from the United States, Canada, Alaska, and Europe. Nevertheless, certain areas were better represented than others. This was due largely to the greater geologic exploration and activity that has taken place in these areas. The extent to which this invalidates the sample is unknown, but it is believed that a sufficient diversification has been obtained to make these effects minor. An attempt was made to use only such sandstones as might be considered typical of their particular locality of origin.

The data for this study were obtained from four sources. Three of these sources were collections of rock specimens which were examined in thin sections, and the fourth was the published literature. The first collection studied was that of the University of Chicago. Although this included about 50 specimens, it nevertheless clearly represented only specific areas of the United States, and many important sandstone-bearing regions were entirely unrepresented. The Chicago Museum of Natural History provided a second collection. From this collection 30 specimens were obtained; these also represented only certain areas of the United States and Europe. Fortunately, this collection represented areas different from those in the University of Chicago's collection. The third and largest collection was the 205 specimens forwarded to the writer by geologists throughout the United States. The author found most of the descriptions in the literature to be in-

[1] Manuscript received December 13, 1948.

complete; hence only those definitive enough to indicate clearly the correct classification of the sandstones were used. Microphotographs, wherever included, proved invaluable in interpreting the descriptions. Many of the relatively recent descriptions were decidedly inferior to descriptions dated back thirty to fifty years.

Even if the varied sources used had yielded a sample of adequate geographic and stratigraphic diversity, many limitations would still remain. First, the author assumed that the relative abundance of the several sandstone types could be obtained by count or tabulation of specific formations and lithologic units, which, however, were of varying thickness and extent. In effect, this assigned equal weight to each specimen irrespective of the amount of sandstone that the particular specimen might represent. Thus a specimen of Dakota sandstone (which represents a formation extending over hundreds of thousands of square miles) carries no more significance than a specimen of the Martinez sandstone (which occurs over a limited area). The actual importance of this factor is unknown.

It would be desirable to use actual volume units in measuring the sandstones, but the problem of arbitrarily selecting and collecting sandstone specimens from specific sections throughout the United States and assigning the appropriate weight to each specimen was not feasible.

The collection of a representative sample of sandstones is further complicated because some distinctive formations are easily delineated and their extent and thickness may be much more precisely known than less distinctive formations. In particular, the graywackes tend to be difficult to measure. These formations commonly occur in thin layers from 2 inches to several feet thick, separated by larger layers of shale or slate. In such cases the graywacke may form one-third to one-fourth the total formation, but the entire unit may still be designated a slate. One of the many examples of this is the Tyler Slate of Michigan.[2] Here, although the total thickness of graywacke may be several thousand feet, the individual beds, not mappable or named formations, are commercially of no importance and therefore not collected, described, or analyzed.

Duplication of specimens is another imperfection in the sample. For example, the final sample used in this study included specimens of the Pottsville sandstone from Alabama, Georgia, and Ohio. All three of these specimens were included because each seemed representative of the sandstones in the particular locality in which the specimen was found, and these localities of origin were widely separated. Where more than one specimen was available from the same formation and the original sources were close together, only one specimen was used.

The assembled data included a total of 277 specimens. Of these, 94 were obtained from the literature and 183 from examination of thin sections. These data included sandstones of all geologic ages, as shown in table 1.

## CLASSIFICATION OF SANDSTONES

The classification used in this study is based on the mineral composition of the detrital grains, and only those specimens in the accepted size range for sandstones were used ($\frac{1}{16}$–2 mm.). Two methods of tabulation were used. In the first classification the sandstones were separated into three series: (1) orthoquartzite series, (2) arkose series, and (3) graywacke

[2] F. J. Pettijohn, personal communication.

series. Krynine has given the following definitions (1948, p. 149):

The quartzite series of sediments can be popularly defined as being made up of "clean sands." It contains one detrital (clastic) end-member, consisting almost exclusively of quartz, occurring generally as medium-sized grains. . . .

The graywacke series of sediments can be popularly defined as a "dark dirty sand" or as a "pepper-and-salt sand" for the lighter-colored but strongly cherty varieties.

Graywackes contain two or more major detrital clastic end-members: grains of quartz, chert, slate, schists, phyllites, etc., and a fine-grained matrix made of finely divided particles of micas (illite, sericite, muscovite, biotite, chlorite). . . .

The arkosic series of sediments can be popularly defined as "ashy, light-gray (or red) dirty sands with much feldspathic materials."

The arkose usually contains one or two major detrital clastic end-members, grains derived from the rapid erosion of a granitic detritus), in some cases mixed with a finer grained matrix of clay minerals (kaolinite-montmorillonite type) and frequently much iron oxide.

TABLE 1

DISTRIBUTION OF SANDSTONE SPECIMENS WITH RESPECT TO GEOLOGIC AGE

| Geologic Age | Observed Specimens | Published Descriptions |
|---|---|---|
| Cenozoic | 10 | 3 |
| Cretaceous | 14 | 15 |
| Jurassic | 14 | 6 |
| Triassic | 14 | 8 |
| Permian | 5 | 4 |
| Pennsylvanian | 32 | 17 |
| Mississippian | 14 | |
| Devonian | 11 | 5 |
| Silurian | 8 | 7 |
| Ordovician | 8 | 10 |
| Cambrian | 18 | 12 |
| Pre-Cambrian | 8 | 10 |

These definitions, while applicable in a general sense, are not precise enough for the separation of certain intermediate-type sandstones, and accordingly the following definitions were used in this paper. The orthoquartzite series includes sandstones with greater than 90 per cent of the detrital constituents made up of quartz grains; they contain, therefore, less than 10 per cent of clay or argillaceous material and also less than 10 per cent of feldspar. The arkose series includes sandstones with greater than 10 per cent of the detrital consituents made up of feldspar and with less than 10 per cent of clay or argillaceous material. The graywacke series includes sandstones containing greater than 10 per cent of clay or argillaceous material.

A second method of tabulation was used to classify the sandstones into more restricted classes. These classes are defined as follows:

1. An orthoquartzite is a sandstone with greater than 90 per cent of the detrital constituents made up of quartz grains.

2. A feldspathic sandstone is a sandstone with feldspar grains making up 10–25 per cent of the total detrital constituents and with the argillaceous components less than 10 per cent.

3. An arkose is a sandstone with feldspar making up greater than 25 per cent of the detrital constituents and with the argillaceous components less than 10 per cent.

4. A graywacke is a sandstone with the argillaceous component making up greater than 25 per cent of the total rock.

5. A subgraywacke is a sandstone with the argillaceous component making up 10–25 per cent of the total rock, plus those sandstones having less than 10 per cent of feldspar or argillaceous material, as well as less than 90 per cent of quartz.

Although the threefold classification has perhaps a wider usage among geologists, this author believes that it is inferior to the fivefold system. The arkose,

graywacke, and orthoquartzite series classification is especially weak in the consideration of the graywacke type. The history of the term "graywacke" from 1808 to the present has been traced by Pettijohn, and he concludes (1944, p. 944) that

> graywacke connotes a type of sandstone marked by (1) large detrital quartz and feldspar ("phenocrysts") set in a (2) prominent to dominant "clay" matrix (and hence absence of infiltration or mineral cement which may on low grade metamorphism (diagenesis) be converted to a chlorite and sericite and partially replaced by carbonate, (3) a dark color, (4) generally tough and well indurated, (5) extreme angularity of the detrital components (microbreccia), (6) presence in smaller or larger quantities of rock fragments, mainly chert, quartzite, slate or phyllite, and (7). . . .

This particular type of sandstone makes up 15 per cent of all the sandstones ex-

TABLE 2

SUMMATION OF THE SANDSTONE TYPES ACCORDING TO GEOLOGIC AGE

| Geologic Age | Observed Sample | | | | | Descriptive Literature | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | Orthoquartzites | Arkoses | Feldspathic Sandstones | Graywackes | Subgraywackes | Orthoquartzites | Arkoses | Feldspathic Sandstones | Graywackes | Subgraywackes |
| Cenozoic | 2 | 2 | 1 | 3 | 3 | 2 | 1 | ..... | ..... | ..... |
| Cretaceous | 9 | ..... | 3 | 2 | 2 | 8 | 1 | ..... | 2 | 4 |
| Jurassic | 5 | 1 | 1 | 3 | 4 | 5 | ..... | 1 | ..... | ..... |
| Triassic | 5 | 3 | 1 | 1 | 3 | 2 | 4 | 2 | ..... | 1 |
| Permian | 3 | ..... | 1 | 1 | 2 | ..... | 2 | 1 | 1 | ..... |
| Pennsylvanian | 12 | ..... | 1 | 4 | 15 | 10 | 1 | ..... | 4 | 2 |
| Mississippian | 11 | ..... | 1 | 1 | 5 | | | | | |
| Devonian | 4 | ..... | ..... | 3 | 6 | 1 | ..... | ..... | 4 | ..... |
| Silurian | 6 | ..... | 1 | 2 | ..... | 2 | 2 | ..... | 1 | 1 |
| Ordovician | 5 | ..... | ..... | 2 | 1 | 6 | ..... | ..... | 2 | 2 |
| Cambrian | 14 | ..... | 4 | 2 | 1 | 5 | 2 | 1 | ..... | 2 |
| Pre-Cambrian | 5 | 4 | ..... | 2 | 1 | 2 | 1 | 2 | 5 | 1 |
| Uncertain | 1 | 2 | 1 | 2 | 3 | ..... | ..... | ..... | ..... | ..... |

TABLE 3

RELATIVE ABUNDANCE OF SANDSTONE TYPES IN PERCENTAGES

| Sandstone Types | Observed Sample | | Descriptive Literature | | Total | |
|---|---|---|---|---|---|---|
| | No. | Percentage | No. | Percentage | No. | Percentage |
| Orthoquartzites | 82 | 45 | 43 | 46 | 125 | 45 |
| Feldspathic sandstones | 15 | 8 | 5 | 5 | 20 | 7 |
| Arkoses | 12 | 7 | 14 | 15 | 26 | 10 |
| Subgraywackes | 46 | 25 | 13 | 14* | 59 | 21* |
| Graywackes | 28 | 15 | 19 | 20 | 47 | 17 |

* The percentage of subgraywackes is probably too low, since many indefinite descriptions of the literature were discarded and, of these, the subgraywackes, being of intermediate type, were probably eliminated disproportionately. The value obtained in the observed sample (25 per cent) is probably more nearly correct.

amined, as shown in table 3. A larger group of sandstones (subgraywacke) has characteristics intermediate between the orthoquartzites and the graywackes. It would seem inaccurate, therefore, to this author, to classify together these two types of sandstones under one general heading (graywacke series) which previously has been defined to include only the smaller group.

TABLE 4

| Rock Types | Per Cent |
|---|---|
| Orthoquartzite | 45 |
| Feldspathic sandstones | 7 |
| Arkoses | 10 |
| Subgraywackes | 21 |
| Graywackes | 17 |

No attempt has been made in this paper to consider unusual or atypical sandstone types; and, of the 185 specimens examined, only 5 per cent did not fit into one of the five "types." These specimens were excluded from the study.

The observed specimens as well as those described from the literature have been classified and assembled in table 2. These data are further summarized in table 3. The relative abundance of each of the sandstone types as shown in this summary is given in table 4.

If these data are computed on the basis of a threefold system of classification, the percentages of each of the types are as shown in table 5.

TABLE 5

| Series | Per Cent |
|---|---|
| Orthoquartzite | 45 |
| Arkose | 17 |
| Graywacke | 38 |

It is quite probable that the values obtained for the orthoquartzites are excessively high because this type has been more extensively collected and described by geologists. The percentage of arkoses and arkose series sandstones are significantly lower than some earlier estimates. Also, as noted in table 3, it is probable that the value shown for the subgraywackes is too low.

The volumetric distribution of the sedimentary rocks (including both conglomerates and sandstones) has also been given by Krynine (1948, p. 156). Considering these apart from siltstones and shales, he finds the percentages of each type shown in table 6.

TABLE 6

| | Per Cent |
|---|---|
| Orthoquartzites | 22½ |
| Low-rank graywacke | 35 |
| High-rank graywacke | 10 |
| Arkose | 32½ |

Krynine also indicates by diagram the average mineral composition of sandstones (p. 151). Measurement of his diagram indicates the average compositions are as shown in table 7. It is important to note that the percentages given by Krynine refer to total composition, i.e., including both detrital and chemical constituents.

TABLE 7

AVERAGE MINERAL COMPOSITION OF COMMON SANDSTONE TYPES IN PERCENTAGES (AFTER KRYNINE)

| Minerals | Arkoses | High-Rank Graywacke | Low-Rank Graywacke | Quartzite |
|---|---|---|---|---|
| Micas and chlorite | 5 | 22 | 22 | 0 |
| Quartz | 48 | 46 | 40 | 79 |
| Feldspars | 25 | 20 | 4 | 3 |
| Clays | 8 | 0 | 5 | 0 |
| Carbonate | 5 | 2 | 7 | 6 |
| Chalcedony | 6 | 7 | 21 | 13 |

If the figures as obtained from the sample used in this study are applied to sandstones in general, it is likely that the individual percentages would vary considerably, but the general order of magnitude should remain.

Because the definitions of each sandstone type (as given earlier in this paper) allow considerable variation within each group, an infinite number of variations was possible; but, in examining the specimens, an effort was made to select those which were typical or ideal examples of the group in question. For example, according to the definitions used, a graywacke might have either a quartz content or a feldspar content, varying anywhere between 0 and 75 per cent; yet, on the basis of actual inspection, the specimens which most clearly conform to the generalized impression conveyed by the term "graywacke" are sandstones having a quartz content of about 25 per cent and a feldspar content of about 20 per cent (plus about 45 per cent argillaceous material). This ideal example is likely to be near an arithmetic mean of all sandstones in this particular type.

Forty specimens, each of which appeared to be a good example of one of the sandstone types, were examined, and the average mineral composition of each type is shown in table 8. The range of each of the more common minerals is also included. It is important to note that the data of table 8 do not include the introduced mineral cements of the sandstones. However, in the case of the graywackes and any of the subgraywackes it is impossible to exclude any particular portion of the rocks on the basis of secondary introduction. Therefore, the entire rock must be included in the measurement of the mineral composition of many of the specimens examined. Utilizing the data of table 7 and table 8, it is possible to estimate the mineral composition of an average sandstone. As shown in table 9, the quartz and feldspar content of the average sandstone has been estimated to be 74 and 8 per cent, respectively.

Using entirely different methods, Clarke has estimated the average quartz and feldspar composition of sandstones to be 66.8 and 11.5 per cent, respectively (1924, p. 33). The general agreement between the values as computed in this paper and those of Clarke indicates that the order of magnitude of the values derived in this paper is correct.

TABLE 8

MINERAL COMPOSITION OF SANDSTONE TYPES IN PERCENTAGES

| Minerals | Orthoquartzites* | | Feldspathic Sandstones† | | Arkoses‡ | | Subgraywackes§ | | Graywackes‖ | |
|---|---|---|---|---|---|---|---|---|---|---|
| | Average | Range | Average | Range | Average | Range | Average | Range | Average | Range |
| Quartz | 95 | 90–99 | 77 | 65–85 | 58 | 40–70 | 75 | 65–83 | 25 | 15–33 |
| Feldspar | 2 | 0–5 | 13 | 10–15 | 27 | 25–30 | 7 | 0–8 | 18 | 2–30 |
| Rock fragments | 1 | 0–4 | 3 | 0–10 | 3 | 0–10 | 5 | 0–6 | 7 | 1–25 |
| Chert | 2 | 0–5 | 5 | 0–10 | 5 | 0–10 | 5 | 1–10 | 4 | 1–10 |
| Argillaceous material | 1 | 0–4 | 2 | 0–5 | 5 | 0–10 | 10 | 2–20 | 46 | 20–60 |

* Includes specimens of the following sandstones: Baraboo, Bass, Cheyenne, Chinle, Graydon, Hepler, Hinkley, Ionian, Jelm, McElmo, Nugget, and Palestine.

† Includes specimens of the following sandstones: Frontier, Goodenough, Hector, Ireland, Kome, LaMotte, and Unkpapa.

‡ Includes specimens of the following sandstones: Hector, Torridonian, and arkosic sandstones from Mount Tom, Massachusetts, Springfield, Massachusetts, and Maryland.

§ Includes specimens of the following sandstones: Bighorn, Dakota, Oswego, Sundance, Third Bradford Sand, and a Cambrian sandstone from the Lake Superior region.

‖ Includes specimens of the following sandstones: Bayard, Catskill, Chico, Franciscan, Monterey, Rensselaer Grit, Upper Slate Member (of Wisconsin), and an Ordovician graywacke from Scotland.

TABLE 9

COMPUTATION OF QUARTZ AND FELDSPAR CONTENT OF THE AVERAGE SANDSTONE

| Types (1) | Percentage of All Sandstones (2) | Percentage of Feldspar (3) | Col. (2)× Col. (3) (4) | Percentage of Quartz (5) | Col. (2)× Col. (5) (6) |
|---|---|---|---|---|---|
| Orthoquartzites | 45 | 2 | 0.0090 | 95 | 0.4275 |
| Feldspathic sandstones | 7 | 13 | 0.0091 | 77 | 0.0539 |
| Arkoses | 10 | 27 | 0.0270 | 58 | 0.0580 |
| Subgraywackes | 21 | 7 | 0.0147 | 75 | 0.1575 |
| Graywackes | 17 | 18 | 0.0306 | 25 | 0.0425 |
| Totals | ..... | ..... | 0.0804 or 8 per cent | ..... | 0.7394 or 74 per cent |

TABLE 10

RELATIONS OF DOMINANT CEMENTS AND THE GEOLOGIC AGE OF SANDSTONES

| Geologic Age | Descriptive Literature | | Observed Specimens | | Literature and Observed Specimens | |
|---|---|---|---|---|---|---|
| | Silica Cement | Carbonate Cement | Silica Cement | Carbonate Cement | Silica Cement | Carbonate Cement |
| Cenozoic | 2 | ..... | 1 | 2 | 3 | 2 |
| Cretaceous | 2 | 5 | 5 | 6 | 7 | 11 |
| Jurassic | 2 | 4 | 6 | 2 | 8 | 6 |
| Triassic | 1 | ..... | 5 | 3 | 6 | 3 |
| Total | 7 | 9 | 17 | 13 | 24 | 22 |
| Permian | 1 | ..... | 3 | 1 | 4 | 1 |
| Pennsylvanian | 6 | 3 | 18 | 7 | 32 | 11 |
| Mississippian | | | 8 | 1 | | |
| Devonian | 1 | ..... | 4 | 2 | 5 | 2 |
| Total | 8 | 3 | 33 | 11 | 41 | 14 |
| Silurian | 4 | ..... | 9 | 2 | 13 | 2 |
| Ordovician | 4 | ..... | 4 | 1 | 8 | 1 |
| Cambrian | 3 | ..... | 15 | 3 | 18 | 3 |
| Pre-Cambrian | 2 | ..... | 6 | 2 | 8 | 2 |
| Total | 13 | ..... | 34 | 8 | 47 | 8 |

### SANDSTONE CEMENTS

The second main objective of this study was the determination of the cementing minerals of the sandstones, with special reference to the ratio of the silica-cemented sandstones to those cemented by carbonate. In cases where two or more different cementing minerals were present, the specimen was disregarded unless one mineral was clearly dominant. Both the rocks described in the literature and the available specimens were classified as to cement. Table 10 presents these data, separated into the geological time units. The variations in the percentages of silica and carbonate cement are summarized in table 11.

The definite increase in the percentage of carbonate-cemented sandstones in the later geologic periods is very evident. This increase may be due to any one of a multitude of factors concerning the deposition and subsequent history of the sandstones, or it may possibly be due to an inherent preference of one particular sandstone type for a particular cement. For example, if the orthoquartzite type sandstone should be present in a larger percentage in the older geologic periods and if this particular type should be predominantly associated with silica cement, this might explain the variation which is found in the cements with respect to age. In an effort to test this possibility we may consider the orthoquartzite type (which has been shown to be the largest single type) more closely.

TABLE 11

| Period | Per Cent Silica | Per Cent Carbonate |
|---|---|---|
| Triassic to Recent | 52 | 48 |
| Devonian to Permian | 75 | 25 |
| Pre-Cambrian to Silurian | 84 | 16 |

First, tables 1, 2, and 3 show that there is no significant variation in the percentage of orthoquartzites with respect to time. The values of table 3 can be summarized as shown in table 12.

TABLE 12

RELATIVE ABUNDANCE OF ORTHOQUARTZITES WITH RESPECT TO GEOLOGIC AGE

| Age | Total No. | No. of Orthoquartzites | Per Cent of Orthoquartzites |
|---|---|---|---|
| Triassic to Recent | 81 | 35 | 47 |
| Devonian to Permian | 108 | 45 | 42 |
| Pre-Cambrian to Silurian | 92 | 48 | 52 |

It can also be shown that there is no variation in any of the other sandstone types of sufficient magnitude to be responsible for the variation in mineral cements which does exist. Furthermore, table 10 shows that 28 per cent of all sandstones examined having a mineral cement are cemented with carbonate. Of a total of 63 orthoquartzites examined, 14, or 22 per cent, showed a carbonate cement. There does not appear to be any significant discrepancy between these two values. Therefore, we may assume that the variation in mineral cements is due not to a variation in the relative amount of each type of sandstone but rather to some physical process present in the history of the sandstone. Just why the carbonate cement is more abundant in the later geologic periods is a question yet unanswered.

This can be further demonstrated by an examination of a sample consisting entirely of one type of sandstone. Table 13 shows the relative abundance of silica and carbonate cements in a group of 77 orthoquartzites, and these data are summarized in table 14.

The values thus summarized from table 13 show marked agreement with those summarized from table 10, and therefore it is again demonstrated that the variation in cement among sandstone studies is independent of the particular type of sandstone.

TABLE 13

DISTRIBUTION OF SILICA AND CARBONATE-CEMENTED ORTHOQUARTZITES WITH RESPECT TO GEOLOGIC AGE

| Period | No. with Silica Cement | No. with Carbonate Cement |
|---|---|---|
| Cenozoic | 1 | 1 |
| Cretaceous | 6 | 2 |
| Jurassic | 1 | 3 |
| Triassic | 3 | 2 |
| Permian | 1 | 1 |
| Pennsylvanian | 10 | 2 |
| Mississippian | 10 | 2 |
| Devonian | 2 | 1 |
| Silurian | 4 | 2 |
| Ordovician | 4 | 1 |
| Cambrian | 13 | 1 |
| Pre-Cambrian | 4 | 0 |

An interesting observation was made on the carbonate-cemented sandstones. Of 44 different specimens examined, almost every one showed some evidence of replacement of detrital quartz grains by carbonate. This process was seen in all stages of development, varying from slight corrosion of some quartz grains to virtually complete replacement of others.

TABLE 14

| Period | Per Cent of Silica | Per Cent of Carbonate |
|---|---|---|
| Triassic to Recent | 63 | 37 |
| Devonian to Permian | 79 | 21 |
| Pre-Cambrian to Silurian | 86 | 14 |

On the basis of the few samples observed, it would appear that this process of replacement or partial replacement is almost universal in the carbonate-bearing sandstones. Should this replacement process continue for a long enough period, it should be possible for a sandstone layer to be completely converted to a limestone. The writer is not aware of any verification of this completed conversion.

## CONCLUSIONS

The following observations and conclusions have been made, based on the examination of the specimens used in this study:

1. On the basis of the detrital mineral components, sandstones may be conveniently separated into five classes, and the relative abundance of these classes is as follows:

| | Per Cent |
|---|---|
| A. Orthoquartzites | 45 |
| B. Feldspathic sandstones | 7 |
| C. Arkoses | 10 |
| D. Subgraywackes | 21[3] |
| E. Graywackes | 17 |

2. A comparison of the carbonate-cemented sandstones with the silica-cemented varieties shows a definite increase in carbonate cement in the later geologic periods.
3. Almost every sandstone specimen showing carbonate cement also displayed partial replacement of the detrital grains by the carbonate.

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[3] As stated in table 7, this value is probably too low, and a better estimation might be 25 per cent.

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# RECENT RESULTS OF INVESTIGATIONS ON THE FELDSPARS[1]

ALEXANDER KÖHLER
Vienna, Austria

ABSTRACT

Careful work with the universal stage on twinned plagioclases has shown that certain optical properties are dependent upon the thermal history of the feldspar. The generally accepted determinative curves for the plagioclases were drawn from feldspars in plutonic rocks and crystalline schists; the volcanic rock plagioclases do not fit these curves. New curves, for high-temperature plagioclases, have been constructed, but are not published here. A discussion of the relation of crystal habit and optical properties of potash feldspars to their genesis is also presented.

No other group of minerals is of so much interest to mineralogists and petrologists as are the feldspars. The feldspars are of such importance as components of the earth's crust that a proper understanding of them is essential for the genesis and classification of rocks. Although scientists of many countries have worked with these minerals, very much more intensive investigation is needed.

This article embodies some recent results bearing on rock genesis and a topic of special importance to petrologists and geologists, namely, the exact determination, with a sensitive and simple method, of the percentage of anorthite in plagioclase as a function of the temperature of crystallization. It will be shown how crystallographic and optical data which seem to be unimportant can be used to interpret the genesis of feldspars.

The first aim in any investigation of plagioclases is the determination of the percentage of An as precisely as possible. Methods for this purpose have been constantly improved. It is possible to get results by using extinction angles in oriented thin sections by means of the "classical methods"; but it is possible to do so only if one takes into consideration the differences, to be discussed later, between plagioclases in plutonic and metamorphic rocks and those in volcanic rocks. The Fedorow universal stage should, however, always be used. No particular technique can be insisted upon here; universal-stage methods are not "universal," and they cannot be used for the investigation of very small crystals, on which extinction methods may be useful. Determinations with the aid of refractive indices, a method which has been elaborated and improved in the United States, shows some factor of uncertainty if one does not watch for the effects to be described later. It is best to use all known methods of determination, checking one against another. One should use the best available method for the material being investigated.

A criticism of the extinction method arises from the fact that plagioclases are not always two-component systems but contain at least three components, the orthoclase content being up to 6 mol per cent. Possibly variation in the amount of orthoclase changes the optical properties and thereby creates an uncertainty in the use of extinction curves. For this reason the extinction methods became discredited, and a preference for the universal-stage method arose. This, however, was a mistake, as will be shown. (It was also

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thought that carnegieite [$NaAlSiO_4$] might be present in solid solution [Washington and Wright, 1910; Barth, 1930]. This should also discredit the use of extinction curves, but later investigations [Dittler and Köhler, 1932; Ernst and Nieland, 1934; and others] showed that practically no carnegieite is taken in the feldspars, although results do not appear to be final on this matter [see, e.g., Lundegardh, 1941].)

If one determines the position of the indicatrix with the aid of the universal stage and relates it to a measured trace (of cleavage or twin plane), as was done by M. Reinhard (1931), M. Berek (1924), and V. Nikitin (1936), it will commonly be noted that the traces of the planes deviate from the position theoretically expected. The determined pole of *M* (010), in particular, at times deviates excessively from the migration curve holding for *M*, for example, in the diagram, table 2, by M. Reinhard (see, e.g., the determinations by G. Paliuc [1932] and others). It was earlier thought that this deviation was due to the potash component and that from this deviation the amount of orthoclase could perhaps be at least estimated (Reinhard, 1931).

There was, at first, no explanation for these deviations. In order to check the above doubtful explanation, the author attempted another approach, and the method is as follows: One measures the indicatrixes of both twins with the universal stage and plots the positions of the main indices, $\alpha$, $\beta$, $\gamma$, and the measured optic axes on the stereographic projection (see the procedure of Reinhard). It is also possible to measure or construct the pole of *M* (it lies on the intersection of the great circles $\alpha\alpha$, $\beta\beta$, and $\gamma\gamma$). Because the optical directions can be measured more exactly than can the crystallographic ones, the construction of the pole should be preferred. In such a diagram it is possible to measure several angles, at least three, namely $\alpha\alpha'$, $\beta\beta'$, and $\gamma\gamma'$ and perhaps some angles between equivalent or inequivalent optical axes of the twins, and to use them for the determination of An content. In the case of the albite law, for example, the angle between the optic axis *A* of the one individual and the axis *B* of the other can be used very successfully (for more detail see Köhler, 1941*a*, *b*).

The extinction methods give only one value that can be used; the universal-stage method also gives the position of but one trace of a plane (rarely two if it is possible to measure poles *M* and *P*). However, our kind of determination leads to the measurement of several angles, all of which should give the same An content if the measurements are correct. The several measurements are mutually consistent and therefore reliable, which is a distinct advance. It is, of course, necessary first to construct the angles between $\alpha$, $\beta$, and $\gamma$ and the optical axes of the two different twins for the known twin laws, using the orientations of "standard plagioclases." This is easily done. In this way one gets for the pericline law the angle $\alpha\alpha^x$, $\beta\beta^x$, $\gamma\gamma^x$, and $AB^x$, $BB^x$; for the Carlsbad law the angle $\alpha_1\alpha_2$, $\beta_1\beta_2$, $\gamma_1\gamma_2$, $A_1A_2$, $B_1B_2$, etc. The same values were used by F. Becke (1906) in his "konoskopischen Zeichentisch-Methoden"; however, he did not use the universal stage and was therefore confined to a few section orientations. The universal stage makes us independent of any specific orientation and enables us to read all necessary angles immediately in the plotted stereogram (see the curves of such angles in Köhler, 1941*a*).

So much may be said in the way of preliminary information. If one now

measures, in the above way, a labradorite with albite twinning from a gabbro and if one plots the determined values of $\alpha$, $\beta$, and $\gamma$ and the optic axes in the stereographic projection (method of Reinhard) and if one plots in the same way the measured position of the pole of $M$, it is then possible to read from the stereogram the following angles (see Köhler, 1941*b*):

$$\alpha\alpha' = 159°, \quad \beta\beta' = 48.5°, \quad \gamma\gamma' = 54°,$$
$$AB' = 44.5°, \quad BB' = 108°.$$

These five angles give, respectively, an anorthite content of

50, 51, 49.5, 51, 50.5 per cent.

Using the position of the pole $M$, one gets in both twins 51 per cent An. The agreement is very good, although admittedly this case was favorable because the twin lamellae were broad and easily measured. The pole of $M$ fits Reinhard's migration curve. If the individuals are not easily measured (and assuming careful work by the microscopist), one will get deviations up to approximately 3° above and below the average. If a measurement is made on an untwinned crystal and one measures a wrong position of the pole of $M$, it would be possible to fit the pole on the migration curve and have it show a value for the An content which appears correct but is not necessarily so. It is important and can be immediately seen in the above example that the albite law holds because the great circle of the pole $M$ is symmetrical to the traces of the axial planes.

If one examines a number of plagioclases of plutonic rocks and crystalline schists in this way, then one will see that within the limit of error the pole of $M$ always lies on the migration curve and the values of the angles all show the same amount of anorthite. However, if one measures a plagioclase of a volcanic rock, quite different relations are found. For instance, using labradorite from an andesite, one will find the following angles and corresponding An contents which, in the first example, are taken from the curves used for the values given above:

$\alpha\alpha' = 158° = 50$ per cent An,

$\beta\beta' = 62° = ?$ An,

$\gamma\gamma' = 67° = 61$ per cent An;

$AB' = 52° = 58$ per cent An,

$BB' = 109° = 56$ per cent An.

It can be seen that the value of the anorthite content varies between 50 and 61 per cent! The high value of 62° for $\beta\beta'$ does not fit on the curve at all. It is therefore impossible to determine the labradorite with this curve. It is not possible to explain this fact as a result of measurements that are in error, since the discrepancy is inherent in the fundamental properties of the feldspar. It is peculiar that in this case the pole of $M$ lies far to the right side of the standard migration curve. Neither can this discrepancy be explained by assuming orthoclase in solid solution; one cannot comprehend why such an amount of orthoclase should be observed only in volcanic rocks. It should be mentioned, moreover, that we have chemical analyses of plagioclases from plutonic rocks with 5–6 mol per cent orthoclase in which the anorthite content is well determinable and the $M$-poles of which lie on the normal migration curve. *The reason therefore seems to be that the optics of volcanic rock-plagioclases (position of the indicatrix) differ fundamentally from those of plutonic rocks.* The high temperature of crystallization of plagioclases in volcanic rocks produces optics which differ measurably from those

of plagioclases of the same chemical composition from crystalline schists or plutonic rocks which are crystallized at lower temperature. As concerns the latter, it can be said without any doubt (as shown below) that the original high-temperature optics changed slowly to the low-temperature optics.

When we know this, the reason for deviation is clarified. It is strange, particularly in the light of our knowledge of the potash feldspars (sanidine-orthoclase-microcline), that the temperature dependency of the optics was not discovered until the sensitive twin optics (Köhler, 1941*a*, *b*) were used. Furthermore, it was due to chance that all plagioclases used for the construction of all the curves were taken from plutonic rocks. Our extinction curves were therefore useless for the determination of volcanic rock plagioclases. It is necessary to establish new curves for such determinations. We have but a few values of optical orientation for such plagioclases (Ernst and Nieland, 1934); we must therefore look for convenient methods of determination. It was thought plausible that plagioclases formed synthetically from a melt should show high-temperature optics. Mixtures were therefore made with 70, 80, 90, and 100 per cent anorthite, and the optical properties of these crystallized synthetic products were measured with the universal stage (see the work of H. Tertsch, 1941*a*, *b*). Another experiment has shown that the position of the indicatrix of low-temperature plagioclases (which were measured before the experiment) was changed to a position measurably different by heating. Tom. F. W. Barth (1931) found incontestable evidence of such a change, which, however, was unjustifiably questioned by C. T. Barber (1936). H. Scholler (1941) studied the temperature dependency of optics in basic and intermediate members in an excellent way. H. Tertsch (1944–45) made a similar study with acid plagioclases, the synthesis of which from the melt is not possible. In this way all curves of the twin optics and all extinction curves for the high-temperature range could be found. These curves are not so precise as the low-temperature curves, but they are exact enough to determine the anorthite content satisfactorily. Turning again to the earlier considered labradorite from the andesite, we find the following values, using the new data:

$$\alpha\alpha' = 158° = 51\text{-}52 \text{ per cent An}$$

$$\beta\beta' = 62° = 54 \text{ per cent An}.$$

$$\gamma\gamma' = 67° = 51.5 \text{ per cent An},$$

$$AB' = 52° = 50 \text{ per cent An},$$

$$BB' = 109° = 52 \text{ per cent An}.$$

This agreement is excellent and is not obtainable with the "low-temperature" curves.

With respect to most plagioclases (andesine to labradorite), the differences between the two ranges of temperature are fairly large and are always clearly distinguishable. It is a disappointing fact, however, that all curves cross at about 80–90 per cent anorthite and that at the albitic end differences are slight. Not many cases may be found, however, in which the distinction cannot be made, as optical measurement of the twins gives several figures, one or the other of which may lead to a decision. For further information see Köhler (1941*a*, *b*) and Tertsch (1941*a*, *b*). The new curves have not as yet been published in any comprehensive form; this will be done in a monograph on the feldspars that the author has ready for publication in the Verlag Springer, Heidelberg, Germany.

What are the advantages of the new type of determination? The more precise nature of mutual checking of angles has already been mentioned. Also, the microscopist, working without the universal stage and using extinction curves, has to use the newly constructed curves because he obtains a value of the anorthite content in the plagioclases of volcanic rocks that is too high by 10–15 per cent through the use of the old curves. An error of this magnitude is unpardonable for the modern petrologist. In no case, therefore, should such plagioclases be determined by other methods, especially not by means of extinction angles. Synthetic plagioclases from the melt can be determined only by means of high-temperature curves; all previous data that indicate that a synthetic pure anorthite, measured optically, shows 100 per cent An are wrong, as all the older methods, when used precisely, really show in this case only 90–92 per cent An.

Another advantage is that one can determine whether the relic in a crystalline schist was originally from a volcanic rock or not (see examples in Köhler, 1941*a*, *b*). We can therefore describe the history of a rock more precisely. It is to be expected, however, that further improvements will be possible that will give us additional information on the past history of plagioclases (reheating in contacts and inclusions, for instance). The author would be grateful for information on measurements of such plagioclases. It would also be interesting to measure the plagioclases of meteorites, something that the author has had no opportunity to do.

When one heats plagioclases of plutonic rocks as mentioned above, one is struck by the fact that they may be brought to the high-temperature form very easily. Crystals that had once been in the high-temperature state are very easily returned to that state; it seems that "relic" structures of the high-temperature form initiate and speed up this process. Material that crystallized at low temperature is very difficult to transform to the high form. The observations of H. Tertsch (1944–45), who found no change in albites from Alpine fissures, with temperatures just below the melting point, but who did find measurable change in albite from Mariupolit treated in the same way, are explainable on the basis of this different behavior.

Similarly important, but until now not so easily described and not so useful in determining genesis as are the results of the research on plagioclases mentioned above, are observations on potassium feldspars, which should be given more consideration than they have received. The type of microcline formed and the type of perthite exsolution that shows different values of the axial angles $2V_\alpha$ give some indication as to the history of the development of the feldspars, and therefore some conclusions can be drawn concerning the history of the rocks, especially the crystalline schists. The crystal habit and the type of twinning is also important; it has been observed after long experience that crystals with elongated *X*-axes or (more rarely) tabular after *M* (010), with the main faces (001), (010), (110), and ($\bar{2}$01) and with subsidiary faces (130), (021), and ($\bar{1}$11), are characteristic of high-temperature formation, possible only in melts. Pegmatitic K-feldspars and adularias, on the other hand, show another habit. At high crystallization temperatures the Carlsbad law is most commonly found, in pegmatites the Baveno law, and adularias for the most part show the Manebach law (associated with the Baveno law), but no Carlsbad twinning. Especially in-

teresting are the Carlsbad twins which overlap and therefore grow somewhat irregularly; the existence of the (2̄01) plane is a necessary condition, however (see Tertsch, 1936; Köhler and Raaz, 1947). Such twins have the (010) plane as the contact plane only in the core, which means that more or less tabular single crystals simultaneously growing in an oriented position became twinned when they were still enclosed in a liquid. When one finds such relations in, for example, granitic gneisses, in which large crystals of this kind quite often are to be found, one can justifiably conclude that these crystals had previously been granite feldspars. When such K-feldspars are twinned according to this law in pegmatites, however, they grow with (010) as the contact plane, and overlapping is not seen because the (2̄01) plane is lacking, the (1̄01) plane being present instead. On the other hand, such "Augen" have often grown in crystalline schists by later infiltration of alkali and silica, which means that it is not necessary to consider them as deformed relics. Hints as to genesis are supplied by such crystallographic observations, but more needs to be done to clarify the issue; for example, the type of microcline cross grating should be considered. This grating depends essentially on the temperature conditions after crystallization and on the tectonic history of the potash feldspar. The amount and kind of perthitic exsolution is connected with all this, and hand in hand with this exsolution can be seen a slow increase in the axial angle $2Va$.

Potash feldspars that have been at high temperatures and rapidly cooled retain the soda content more or less isomorphously, microcline grating is not visible, and the axial angle remains small. With slower cooling, perthitic exsolution appears, and the grating is obvious (see granite feldspars). The size of $2Va$ ranges between 65° and 75°. The potash feldspars that originally certainly crystallized in the monoclinic form (Köhler, 1948) change to a triclinic modification; this change, being aided by tectonic forces which, with increased intensity, increases the "coarseness" of the exsolution of albite and produces microcline grating. It is always quite obvious that the grating is most distinct in the neighborhood of the broad albite "stringers," as in these places the exsolution of the albite destroys the originally monoclinic lattice, and the newly crystallized K-feldspar assumes the triclinic modification which is now the stable one. The crystal, by twinning polysynthetically according to the albite and pericline laws, attempts to retain its past monoclinic habit (Köhler and Raaz, 1947).

When the original rock (granite, for example) is very strongly changed to a crystalline schist (granite gneiss), all the albite will be exsolved and will form independent albite crystals in the gneiss, and the K-feldspar nearly free of soda will have very coarse grating structure. In addition, the axial angle (around $a$) of the K-feldspar will be $>80°$ and, in fact, $>90°$ more often than has been realized (optically positive "isomicrocline").

It would therefore be possible, using the phenomena mentioned above, to obtain some insight into magmatic and crystalloblastic formation if it were not for the fact that pegmatitic liquids and solutions and formation of feldspars in deep zones of mountain roots cloud the picture. The perthitic K-feldspar from a pegmatite looks very much like strongly exsolved granite feldspar, as regards grating and exsolution structures. If no Carlsbad twinning is present to give one a hint as to the origin of the feldspar, wrong

conclusions might be drawn. In such cases field observations may tell more than microscopic investigation of the K-feldspars, and it is necessary to use observation on the plagioclases. If the typical polysynthetic albite twins in the plagioclases are lacking, we have an indication that the crystals were formed at low temperature (for example, granite gneisses of the upper Katazone, such as in the central Alps). If there are multiple repetitions of the lamellae, it is suggestive of high temperature at an earlier time or that the plagioclases came from pegmatites. It can be said without doubt that the multiple repetition of albite lamellae is not a primary result of crystallization but is secondarily induced in the plagioclase. We (Köhler and Raaz, 1947) think that the modification of the plagioclases formed originally from hot melts was a monoclinic one or very nearly monoclinic, or it may be closer to being monoclinic than the plagioclases formed at low temperatures. As K-feldspar with lowering of the temperature is transformed to a finely twinned arrangement, so plagioclases may show a similar effect when the temperature is lowered and mechanical effects come into play. I have tried (1948) to stimulate petrogenetic evaluation of these observations, but the subject still appears somewhat delicate. In a paper by C. Exner (1949) it is shown that the interplay of the influences that change K-feldspar *and* plagioclase gives us some interesting insight into the whole problem.

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# DIE HÄUFIGKEIT DER EDELGASE AUF DER ERDE UND IM KOSMOS[1]

HANS E. SUESS
Institut für physikalische Chemie der Universität Hamburg

## ENGLISH SUMMARY

The cosmic abundance of the rare gases is now known with sufficient accuracy to allow a quantitative comparison with the corresponding data for the earth's atmosphere. It seems safe to assume that the cosmic abundance of neon is closely approximated by Unsöld's value forthe neon content of the Bo star, $\tau$ Scorpii. The values for the other rare gases have been determined in a way previously described by the author, viz., by interpolation of the relative values for the different nuclear species in "smoothed" abundance curves. These values, as well as the neon value, may be considered to be correct within a factor of 2. Argon 40 and the two helium isotopes are omitted from consideration because their atmospheric concentration is of radiogenic origin. The comparison of these values with those for the composition of the earth and its atmosphere show that the relative abundance of xenon (the ratio of the number of xenon atoms to the number of silicon atoms) is about $10^7$ times smaller on the earth than in the universe, whereas for neon this figure exceeds $10^{11}$. A separation of neon from xenon, shifting the ratio of concentrations by a factor of more than $10^4$, has obviously taken place during the process of the atmosphere's evolution. This cannot be explained by chemical absorption or solution processes but can be understood by assuming selective diffusion from the earth's gravitational field during a limited epoch of evolution.

In greater detail, it is found that the ratios of terrestrial and cosmic abundances of Ne, $A^{36}$ plus $A^{38}$, Kr, and Xe, plotted against atomic weight, lie close to a curve, which is given by the following equation:

$$-\log_{10}\frac{N_{ter}}{N_{sol}} = 10 \times e^{-0.045M/m_1} + 7.1 \ ,$$

in which $N_{ter}$ = atoms of rare gas present in the earth's atmosphere per 100 atoms of Si present in the earth; $N_{sol}$ = atoms of rare gas in the universe per 100 atoms of Si; $M$ = atom weight of rare gas; and $m_1$ = atomic mass unit.

The form of this function corresponds to what one would expect to result from selective diffusion. Further discussion of the equation shows that the remarkably small value of the exponent makes it impossible to explain the selective escape from the earth's field of gravitation as a consequence of high temperature only, but that the small value can be understood only if the short period of the earth's rotation at the time of evolution is taken into account. The equation indicates that approximately $10^7$ parts to 1 of the rare gases originally present may have escaped from terrestrial matter without undergoing separation, leaving only the small residual fraction to be involved in the separation process.

Attention is drawn to the fact that, according to the given equation, a marked difference should exist between the isotopic composition of atmospheric and that of cosmic rare gases. An investigation of the isotopic composition of Ne and Ar obtained from meteorites, rocks, or natural gases containing an excessive amount of neon might lead to a final decision on the correctness of the assumption suggested.

In den letzten Jahren ist, vor allem durch die grundlegenden Arbeiten des Mineralogen V. M. Goldschmidt (1938) und des Astronomen A. Unsöld (1944) die alte Auffassung in überzeugender Weise bestätigt worden dass sich die Erde und die anderen Planeten, ebenso wie die Meteorite aus einer gasförmigen Urmaterie gebildet haben, die dieselbe chemische Zusammensetzung besass, wie heute die Sonne und auch die Mehrzahl der anderen Fixsterne. Der Verfasser selbst hat (1946) eine Reihe von Argumenten geltend machen können, durch die gezeigt wurde, dass diese überraschend gleichförmige chemische Zusammensetzung der kosmischen Materie im Wesentlichen einer Häufigkeitsverteilung entspricht, mit der sich die Elemente bei einem Entstehungsprozess gebildet haben.

Das Mengenverhältnis aller jener Elemente, die bei einer Temperatur von etwa 2000° K zur Kondensation gelangen, scheint in der Gesamtmenge der

[1] Manuscript received February 23, 1949.

Erde und in der Gesamtmenge der Meteoriten mit dem auf der Sonne in engen Grenzen übereinzustimmen; diejenigen Elemente jedoch, die bei der genannten Temperatur flüchtig sind oder in einem entsprechenden Medium flüchtige Verbindungen bilden, wie Wasserstoff, Kohlenstoff und Stickstoff, sowie die Edelgase, sind mengenmässig auf der Erde und in den Meteoriten zu einem wesentlich geringeren Anteil vertreten, als auf der Sonne. Es ist klar, dass diese Abtrennung der flüchtigen Anteile der Urmaterie stattgefunden haben muss, nachdem der überwiegende Teil der Erdmaterie bereits in kondensiertem Zustand vorhanden war, denn ein Prozess, der in der Gasphase etwa zu einer praktisch quantitativen Trennung des Xenon vom Jod oder Cäsium geführt haben könnte, ist völlig undenkbar.

In den letzten Jahren ist es gelungen, verhältnismässig präzise Angaben über die kosmischen Edelgashäufigkeiten zu machen (vergl. Suess, 1949). Diese Fortschritte gaben die Veranlassung dafür, das bereits von Russell und Menzel (1933) eingehend behandelte Problem der irdischen Edelgashäufigkeiten neu aufzugreifen. Auch besteht heute im Wesentlichen Klarheit über den Ablauf der Kernprozesse, durch die die beiden Heliumisotope $He^3$ und $He^4$, sowie das Argonisotop $A^{40}$ laufend nachgebildet werden. Der Vorgang der Bildung von $He^3$ wurde von Libby (1946) diskutiert; der experimentelle Nachweis für die bereits früher (Weizsäcker, 1937; vergl. auch z B. Bleuler u. Gabriel, 1947; Graf, 1948; Suess, 1948) vermutete Bildung von $A^{40}$ durch K-Umwandlung aus $K^{40}$ ist Aldrich und Nier (1948) durch massenspektroskopische Untersuchung von Argon aus kaliumreichen Gesteinen gelungen. Helium, und zwar beide Isotope wandern ständig aus der Erdatmosphäre ab, denn die Atmosphäre enthält nach Goldschmidt (1938) nur etwa 1 % des $He^4$, das in den letzten $2 \cdot 10^9$ Jahren radiogen aus Thorium und Uran gebildet und aus dem Eruptivgestein im Kreislauf der Verwitterungsvorgänge in Freiheit gesetzt wurde.

Die in Tab. 1 angegebenen Werte beziehen sich auf die anderen, nichtradiogenen Edelgase, die Werte für Argon somit nur auf die Isotope der Masse 36 und 38, denn der ursprüngliche Anteil von $A^{40}$ im kosmischen Argon ist sicherlich kleiner als 1 %, wie eine Abschätzung mit Hilfe der Häufigkeitsregeln (Suess, 1946) zeigt. Für Neon ist die kosmische

TABELLE 1

DEKADISCHE LOGARITHMEN DER HÄUFIGKEITEN SOWIE DES HÄUFIGKEITSVERHÄLTNISSES DER NICHTRADIOGENEN EDELGASE IN DER ERDATMOSPHÄRE UND IM KOSMOS*

| (1) El. | (2) $M$ | (3) Vol.-% der Atmosphäre | (4) $\log H_{tar}$ | (5) $\log H_{sol}$ | (6) $\log \frac{H_{tar}}{H_{sol}}$ | (7) Vorhandene % d. Restm. |
|---|---|---|---|---|---|---|
| Ne | 20,2 | 0,00161 | − 8,00 | +3,26 | −11,26 | 0,007 |
| Ar | 36,3 | 0,00345 | − 7,67 | +1,20 | − 8,87 | 2,1 |
| Kr | 83,8 | 0,000108 | − 9,17 | −1,82 | − 7,35 | 55 |
| X | 131,3 | 0,000008 | −10,31 | −3,17 | − 7,14 | 91 |

* Die Häufigkeitswerte sind bezogen auf die Gesamthäufigkeit des Si, die zur Normierung gleich 100 gesetzt ist.

Häufigkeit von Unsöld (1941) durch quantitative Spektralanalyse des Bo-Sternes $\tau$ Scorpii direkt bestimmt worden. Dieser Stern besitzt, wie Unsöld durch Vergleich der Häufigkeitswerte von sechs anderen Elementen zeigen konnte, sicher dieselbe chemische Zusammensetzung, wie die Sonne. Für die übrigen Edelgase sind vom Verfasser (1949) Häufigkeitswerte angegeben worden, die durch Anwendung von Regeln für die kosmischen Kernhäufigkeiten nach einem Interpolationsverfahren gewonnen worden waren. Bei diesem Verfahren werden zur Interpolation die Häufigkeitswerte der einzelnen Isotopen gesondert herangezogen und Werte erhalten, deren Fehler kaum einen Faktor 2 erreichen dürfte. Die Logarithmen dieser Werte, wie üblich bezogen auf die kosmische Häufigkeit von Silizium gleich 100, sind in Spalte 5 der Tabelle angegeben. In Spalte 3 ist der Gehalt der atmosphärischen Luft an Edelgasen in Volumprozent, in Spalte 4 der Logarithmus der Gesamtmenge Edelgas in der Luft in Atomprozent der Gesamtmenge Silizium im Erdball angegeben, wobei die Siliziummenge zu 15% der Erdmasse angenommen wurde. In Spalte 6 sind die Differenzen der Werte von Spalte 4 und 5, d.h. die Logarithmen der Bruchteile der Edelgasmengen angegeben, die in der Atmosphäre zurückgeblieben sind. Diese Werte sind in Abb. 1 als Funktion des Atomgewichtes dargestellt.

Dass der Edelgasgehalt der Erdatmosphäre ein gutes Mass für die relativen Mengen der Edelgase an der Erdoberfläche ist, lässt sich durch eine genaue Diskussion der Löslichkeiten und Absorbtionskoeffizienten zeigen. Man sieht dann, dass selbst von dem am leichtesten löslichen Xenon nur wenige Prozent der in der Atmosphäre enthaltenen Menge in den Ozeanen gelöst sind, und dass auch die Adsorbtion an den Schneemengen der Polkappen oder an trockenen Wüstensanden gegenüber der Menge freien Xenons nicht ins Gewicht fällt.[2]

Die Zahlenwerte der Spalte 6 bringen zum Ausdruck, dass das Xenon relativ zum Silizium auf der Erde um rund 7 Zehnerpotenzen, Neon jedoch um über 11 Zehnerpotenzen seltener ist, als auf der Sonne. Es ist somit, im Verhältnis, mehr als 10 000 mal soviel Xenon auf der Erde verblieben, als Neon. Dieser Befund lässt sofort erkennen, dass die Edelgase in ihrem gegenwärtigen Mengenverhältnis nicht durch vulkanische Exhalationen oder aus verwitternden Gesteinen abgegeben worden sind, denn Lösungs- oder Adsorbtionsvorgänge können selbst bei so niederen Temperaturen, wie sie heute auf der Erdoberfläche herrschen, eine Trennung des Neon vom Xenon höchstens um etwa einen Faktor 30 hervorrufen. So beträgt z.B. das Verhältnis der Löslichkeiten von Xenon und Neon in Wasser bei 0° C: 11, 4, bei 50° C: 5, 6 (S. Valentiner, 1927). Die Löslichkeit dieser Edelgase im magmatischen Schmelzfluss dürfte sich wegen der hohen Temperatur kaum um mehr als einen Faktor 2 unterscheiden. Es ist völlig ausgeschlossen, dass Vorgänge einer selektiven Adsorbtion oder Lösung einen Trennfaktor der Grössenordnung $10^4$ hervorrufen können.

Das gegenwärtige Mengenverhältnis der Edelgase in der Erdatmosphäre lässt sich kaum anders deuten, als durch die Annahme, dass während einer bestimmten Epoche Bedingungen vorlagen, unter denen eine selektive Diffusion aus dem Schwerefeld stattfinden konnte. Die pro Zeitinterwall $dt$ aus dem Schwerefeld eines

[2] Herrn stud. B. Franck danke ich für einige ergänzende Versuche, die er in diesem Zusammenhang zur Bestimmung der Xenonadsorbtion an Quarzsand ausgeführt hat.

Planeten entweichende Menge-$dN$ eines Gases ist eine Funktion der Masse $M$ der Gasteilchen, die näherungsweise gegeben ist durch den Ausdruck (näheres z.B. bei J. H. Jeans, 1923, od. R. H. Fowler, 1929):

$$-dN = N_t \cdot A \cdot e^{-Mgr/RT}. \quad (1)$$

Hierin bedeutet: $N$ eine Anzahl von Molekeln, $N_t$ die zur Zeit $t$ vorhandene Anzahl und $M$ das Molgewicht des betrachteten Gases, $g$ die Schwerebeschleunigung in der Entweichungszone und $r$ deren Abstand vom Gravitationszentrum, ferner $R$ die Gaskonstante, $T$ die absolute Temperatur und $A$ eine im Wesentlichen von $M$ unabhängige Konstante. Für die nach der Zeit $t$ noch im Schwerefeld des Planeten vorhandene

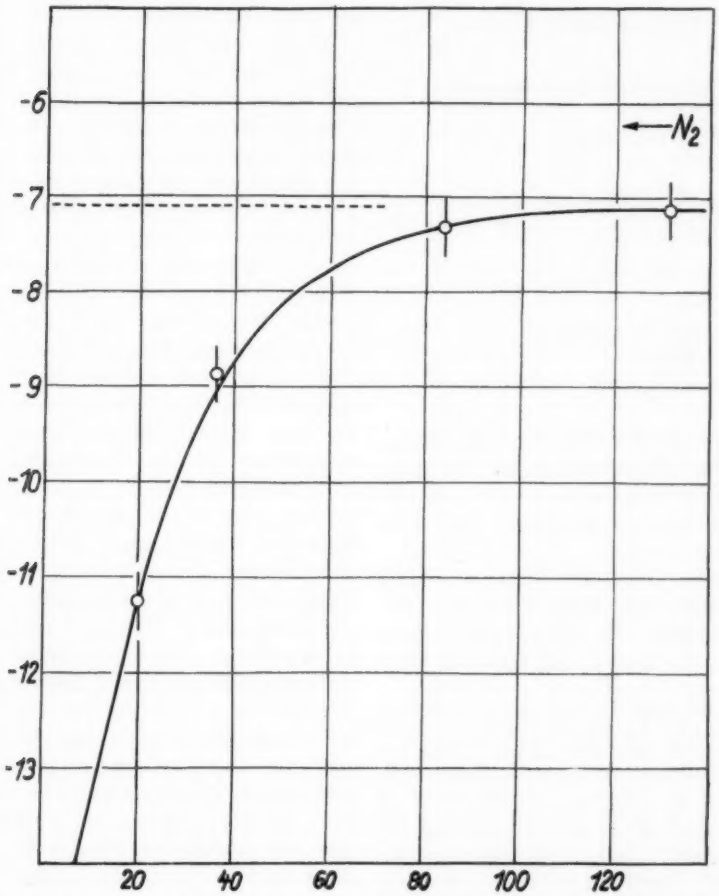


ABB. 1.—*Abszisse:* Atomgewicht $M$. *Ordinate:* Dekadischer Logaritmus des Verhältnis der Häufigkeiten der Edelgase in der Erdatmosphäre und im Kosmos ($\log N_{ter}/N_{sol}$), bezogen auf die Häufigkeit des Siliziums. Die ausgezogene Kurve entspricht der Funktion:

$$-\log N_{ter}/N_{sol} = 10 \cdot e^{-0{,}045\, M/m_1} + 7{,}1\,.$$

Menge eines Gases mit dem Molekulargewicht $M$ gilt somit:[3]

$$-lnN_t = t \cdot A \cdot e^{-Mgr/RT} - lnN_0 \,. \quad (2)$$

Die gegenwärtige Häufigkeitsverteilung der Edelgase in der Atmosphäre entspricht nun in der Tat einer solchen Funktion, und zwar entspricht die in Abb. 1 eingezeichnete Kurve der Beziehung:[4]

$$-ln\,\frac{N_{ter}}{N_{sol}} = 23 \cdot e^{-0,045M/m_1} + 16.3\,, \quad (3)$$

oder in dekadischen Logarithmen:

$$-\log\frac{N_{ter}}{N_{sol}} = 10 \cdot e^{-0,045M/m_1} + 7.1\,. \quad (4)$$

Für $N_t = N_{ter}$ ergibt ein Vergleich von Gl. (2) mit (3) bezw. (4) die folgenden Werte:

$$N_0 = 10^{-7,1} \cdot N_{sol}\,,$$

$$\frac{Mgr}{RT} = 0,045\,\frac{M}{m_1}\,,$$

$$t \cdot A = 23\,.$$

Den genauen Zahlenwerten soll wegen der vielen Unsicherheiten kein besonderes Gewicht beigelegt, sondern es sollen nur die Grössenordnungen diskutiert werden.

Vor allem ist der ausserordentlich kleine Wert des Exponenten $Mgr/RT$ von Interesse. Für den gegenwärtigen Zustand der Erde erhält man mit einer Atmosphärenhöhe von rund 500 km (vergl. R. H. Fowler, 1929, S. 396) und einer Temperatur von rund 300° K für den Exponenten $Mgr/RT \approx 30\ M/m_1$. Merkliche Bruchteile schwererer Gase ($M/m_1 > 10$) können unter diesen Bedingungen in den Zeiträumen seit der Bildung der Erde nicht entwichen sein. Bei höherer Temperatur ist der Wert des Exponenten kleiner, jedoch wird selbst bei einer Temperatur von 3000° K und einer dieser Temperatur entsprechenden Atmosphärenhöhe von rund 15 000 km der Exponent $Mgr/RT \sim 1\ M/m_1$ (Näheres bei Russell u. Menzel, 1933). Ein hoher Atmosphärendruck vergrössert, wie man leicht sieht, die Atmosphärenhöhe nicht massgeblich. Dass sich das Entweichen der Edelgase bei noch höherer Temperatur vollzogen hat, ist nicht möglich, da dann die Silikate und andere Substanzen der Erdoberfläche bereits in hohem Masse flüchtig sind.

Es bleibt somit nur die Möglichkeit, dass das Gravitationspotential bezw. die effektive Schwerebeschleunigung während einer gewissen Zeitdauer kleiner war, als heute. Eine derartige Annahme scheint zunächst für die "Planetesimalhypothese" zu sprechen, nach der sich die Erde aus kleineren Körpern gebildet haben soll. In der Tat kann die Atmosphäre eines Himmelskörpers von etwa 1 bis 0, 1 % der Erdmasse unter geeigneten Bedingungen einen Zustand durchlaufen, in dem die Häufigkeitsverteilung der Edelgase ihrem gegenwärtigen irdischen Häufigkeitsverhältnis entsprechen würde. Dieser Zustand wird jedoch, wie eine Abschätzung der Geschwindigkeit der Abdiffusion (Näheres z.B. bei E. A. Milne, 1923) zeigt, bereits nach so kurzer Zeit durchlaufen, dass man diesen Körpern vom Zeitpunkt ihrer Bildung aus Solarmaterie bis zu ihrer Vereinigung zum Erdkörper nur eine Lebensdauer

[3] Derselbe Funktionsverlauf gilt auch für den Fall vollständiger Ionisation der Gasschicht, aus der das Entweichen stattfindet (Näheres hierüber bei W. Kuhn, 1948). Bei vollständiger einfacher Ionisation tritt in Gl. (1) und (2) anstelle von $M$ im Exponenten die Grösse $M - \overline{m}/2$, wobei $\overline{m}$ das mittlere Molgewicht der die Atmosphäre bildenden Gase bedeutet. In diesem Fall erfolgt somit das Entweichen lediglich um den für alle in der Atmosphäre enthaltenen Gase gleichen Faktor $\exp(\overline{m}gr/RT)$ rascher.

[4] Aus dimensionellen Gründen ist hier das Atomgewicht $M$ durch die Atomgewichtseinheit $m_1$ dividiert.

von wenigen Stunden oder allenfalls Tagen zubilligen könnte.

Diese Schwierigkeit sowie die bekannten allgemeinen Einwände gegen die "Planetesimaltheorie" lassen es plausibler erscheinen, zur Deutung des kleinen Wertes des Exponenten die hohe Rotationsgeschwindigkeit der Erde zur Zeit ihrer Bildung in Betracht zu ziehen. Wie H. Jeffreys (1924) gezeigt hat, verlangsamt die Erde ständig unter dem Einfluss der durch den Mond verursachten Gezeitenreibung ihre Rotation. Sie hat sich in der ersten Zeit nach ihrer Entstehung sicherlich in weniger als 6 Stunden um ihre Achse gedreht. Für eine wie ein starrer Körper rotierende Atmosphäre verschwindet bei dieser Rotationsgeschwindigkeit die effektive Schwerebeschleunigung am Äquator bereits in einer Höhe von rund 10 000 km. Man kann leicht abschätzen, dass eine Atmosphäre, deren Gasmenge der gegenwärtigen entspricht, bereits bei Temperaturen von etwa 2000° K in diese Gegend hinaufreicht.

Es sei deshalb das folgende Bild vom Vorgang der Edelgasabwanderung vorgeschlagen, das sich vermutlich am ehesten in ein umfassenderes Bild der Planetenentstehung einordnen dürfte:[5] In einer ersten Phase nach der Kondensation der Erdmaterie (vergl. A. Eucken, 1944) ist der Hauptteil der in der ursprünglichen Solarmaterie enthalten gewesenen Edelgase, und zwar das rund $10^7$-fache des zurückbleibenden Restes infolge der hohen Rotationsgeschwindigkeit, vermutlich auch zum Teil noch als dynamischer Bestandteil einer das ganze Planetensystem umfassenden turbulenten Gasmasse (im Sinne v. Weizsäckers, 1944), konvektiv, ohne das Auftreten von Separationseffekten von dem kondensierten Anteil der Erdmaterie abgetrennt worden. In einer sich anschliessenden zweiten Phase wurde bei abnehmender Atmosphärenhöhe und Rotationsgeschwindigkeit ein Stadium erreicht, in dem die Gasatome das Schwerefeld der Erde am Äquator nur mehr über einen, wenn auch kleinen Potentialwall verlassen konnten. Die zu Beginn dieses Stadiums vorhanden gewesene Neonmenge würde unter heutigen Bedingungen einem Partialdruck von etwa 200 mm Hg entsprechen. Der Gesamtvorgang würde einer allgemeinen Eigenschaft rotierender kosmischer Gasmassen entsprechen, die nach v. Weizsäcker (1948) sich stets in einen Kern und einen ins Unendliche entweichenden Teil auflösen.

In dem Stadium des Überganges der Abwanderung der Edelgase durch Konvektion in Abdiffusion kann sich bei abnehmender Temperatur und Atmosphärenhöhe die Separation der Edelgase vollzogen haben. Es ist kaum möglich, quantitativere Aussagen über den Verlauf des Prozesses zu gewinnen, da alle massgebenden Parameter, wie Erdrotation, Temperatur und Atmosphärenmenge, sowie ihre zeitlichen Änderungen in weiten Grenzen unbestimmt sind. Es scheint jedoch unmittelbar plausibel, dass ein derartiger Prozess näherungsweise eine Verteilung nach Gl. (3) ergeben kann.

Abschliessend sei noch darauf hingewiesen, dass eine derartige nach dem Atomgewicht differenzierte Abdiffusion der Edelgase aus dem Schwerefeld der Erde zwangsläufig auch zu einer Verschiebung der isotopischen Zusammensetzung der zurückbleibenden Gase geführt haben müsste, da Gl. (3) natürlich auch für Isotope Atomarten gilt. Man wird daher vermuten, dass die Edelgase der Erdatmosphäre etwas reicher an ihren schwereren Isotopen sind, als im

[5] Vergl. Zeitschr. f. Elektrochemie (1949), Bd. 53, S. 237.

Kosmos. Im Falle des Neons haftet dieser Überlegung freilich noch eine zusätzliche Unsicherheit an: Es wäre möglich, dass die Erdatmosphäre unmittelbar nach der Edelgasabwanderung noch ärmer an Neon war, als heute und einen Teil des Neon durch nachträgliche Aufnahme geringer Mengen von unentmischter Solarmaterie erhalten hat. In Abb. 1 würde dann der Punkt für den ursprünglichen Wert des Neon noch tiefer liegen und der gegenwärtige Wert würde sich nur zufällig in den glatten Funktionsverlauf einordnen. Nimmt man jedoch an, dass alles atmosphärische Neon am Entmischungsprozess beteiligt war, dann erhält man nach Gl. (4) die in Tab. 2 angegebene isotopische Zusammensetzung des kosmischen Neon (s. Tab. 2).

TABELLE 2

| | Massenzahl | | |
|---|---|---|---|
| | 20 (%) | 21 (%) | 22 (%) |
| Atmospärisches Ne..... | 90,51 | 0,28 | 9,21 |
| Kosmisches Ne (berechn.)... | 93,4 | 0,2 | 6,4 |

Für die anderen Edelgase gilt die für Neon gemachte Einschränkung jedoch nicht. Für das Verhältnis $A^{36}:A^{38}$ müsste sich noch ein deutlicher Unterschied zwischen atmosphärischem und kosmischem Argon feststellen lassen. In Tab. 3 sind die Daten für die isotopische Zusammensetzung von Argon angegeben. Die Werte der zweiten Zeile erhält man, wenn man von dem aus $K^{40}$ gebildetem $A^{40}$ absieht und nur den Gehalt an primär vorhanden gewesenem $A^{40}$ berücksichtigt, der durch Interpolation aus den Daten für die kosmische Häufigkeit von $S^{36}$ und $Ca^{44}$ abgeschätzt werden kann (Suess, 1946). Die Werte der dritten Zeile sind wie oben nach Gl. (4) berechnet.

In der Literatur finden sich mehrfach Angaben über einen auffallend hohen Neongehalt von natürlichen Gasen und Gesteinen (A. Gautier, 1909; W. Ramsay, 1912; Lord Rayleigh, 1938). Es ist möglich, dass in manchen Tiefengesteinen noch Edelgasmengen enthalten sind, die nicht oder nur teilweise den Separationsprozess der zweiten Phase der Edelgasabwanderung durchgemacht haben und deswegen vorwiegend aus Neon bestehen. Die Kenntnis der isotopischen Zusammensetzung von Neon und Argon aus solchen natürlichen Gasen und Gesteinen könnte eine Entscheidung in diesen Fragen bringen. Es ist auch mit der Möglichkeit zu rechnen, dass die beobachteten Schwankungen in der isotopischen Zusammensetzung des Kohlenstoffs (Näheres bei K. Rankama, 1948) zum Teil durch die Trenneffekte bei der Abwanderung der flüchtigen Kohlenstoffverbindungen aus dem Erdschwerefeld hervorgerufen wurden. Hier sind jedoch die Verhältnisse durch die Lage der Isotopenaustauschgleichgewichte besonders verwickelt. Nach Drucklegung der vorliegenden Arbeit ist dem Verfasser der Artikel von Harrison Brown "Rare gases and the formation of the earth's

TABELLE 3

| | Massenzahl | | |
|---|---|---|---|
| | 36 (%) | 38 (%) | 40 (%) |
| Atmosphärisches A.... | 0,306 | 0,061 | 99,6 |
| Ds. ohne radiogenes A..... | 83,0 | 16,5 | ~ 0,5 |
| Kosmisches A (berechn.)... | 85,6 | 14,1 | ~ 0,4 |

atmosphere"[6] bekannt geworden, dem gleichfalls ein Vergleich der kosmischen und terrestrischen Edelgashäufigkeiten zugrundegelegt ist. Die hier dargelegten, und die von Harrison Brown ausgeführten Überlegungen beruhen praktisch auf den selben Grundannahmen und ergänzen sich in vielerlei Hinsicht.

Den Herren P. Harteck und H. Jensen danke ich herzlich für wertvolle Diskussionen über dieses Thema. Es sei mir auch gestattet, der Geological Society of America an dieser Stelle zu danken für die Vermittlung privater Unterstützungen im vergangenen Jahre, durch die mir damals die Weiterführung wissenchaftlicher Arbeiten ermöglicht wurde.

[6] Aus: "The atmospheres of the earth and planets" Ed. G. P. Kuiper, Chicago, University of Chicago Press, 1949, 260 pp.

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# A NOTE ON THE GEOCHEMISTRY OF THALLIUM[1]

KALERVO RANKAMA
University of Chicago[2]

ABSTRACT

The manner of occurrence of thallium in the upper lithosphere is discussed. The value 0.6 g/ton is tentatively adopted for the abundance of thallium in igneous rocks. The presence of notable variations in the Rb/Tl ratio in igneous rocks is established. The geochemical coherence of the pair Rb-Tl is discussed and compared with the coherence in the pairs Ga-Al, Cb-Ta, and Zr-Hf, and in the group of lanthanum and the lanthanides. It is concluded that zirconium and hafnium probably form a more coherent pair than do rubidium and thallium.

## INTRODUCTION

So far, thallium has been and partly still remains one of the geochemically more imperfectly known elements. The most recent contributions to the geochemistry of thallium consist of a series of papers by Ahrens (1945*a*, *b*; 1947*a*, *b*; 1949*a*, *b*) which have shed much welcome light on the mineralogical geochemistry of this metal. Ahrens emphasized the peculiar chemical character of thallium: the unipositive thallium ion behaves chemically much like the ions of the alkali metals of higher atomic weight. He analyzed spectrochemically 167 mineral samples for their content of $Rb_2O$ and $Tl_2O$. This material consisted of primary and hydrothermal pegmatitic microclines, amazonite, plagioclase, muscovite, phlogopite, lepidolite, zinnwaldite, biotite, pollucite, rhodizite, and cesian beryl. The results revealed a very close and unique association of rubidium and thallous ("alkali-metal") thallium in potassium minerals. The $Rb_2O/Tl_2O$ ratio was found to lie between 12 and 650; the mean ratio was 100. Ahrens found no shift in the Rb/Tl ratio "throughout the selective crystallization of minerals." Therefore, he concluded, rubidium and thallous thallium comprise a pair of elements which are remarkable for the constancy of their relative proportions; the relationship apparently is independent of the mineral type; it holds over all ranges of concentration; moreover, if the analyses are confined to a single area or source, the $Rb_2O/Tl_2O$ ratio remains almost exactly constant within the limits of experimental error and even approaches an isotopic constancy. Consequently, "alkali-metal" thallium and rubidium are very closely associated geochemically in the upper lithosphere, and it is possible that their relationship is closer than the relationship between any other pair of elements, with the possible exception of zirconium and hafnium and some of the rare-earth metals. Moreover, Ahrens used the newly established $Rb_2O/Tl_2O$ ratio to calculate a new value for the abundance of thallium in the upper lithosphere because he thought that the close association of thallium and rubidium in minerals would make possible a reasonable estimate of the abundance of thallium, derived from the abundance of rubidium. For the abundance of rubidium in igneous rocks Ahrens used Goldschmidt's (1937*a*, *b*) value of 300 g/ton (actually, 310 g/ton), which he thought to be reasonably ac-

[1] Manuscript received January 31, 1949.

[2] On leave from the University of Helsinki.

curate; and for the abundance of thallium in igneous rocks he adopted the value 3 g/ton, which is ten times as high as the old value, or 0.3 g/ton, of Goldschmidt (1937*a*, *b*). It may be added here that the previous abundance values of thallium in the upper lithosphere calculated by Noddack and Noddack (1930, 1934) are still lower, viz., 0.1 and 0.01 g/ton, respectively. Ahrens claimed that Goldschmidt's value is based on relatively scant data and found that the old value was evidently too low. Finally, Ahrens pointed out that the new value refers to the "alkali-metal" thallium; but, because the content of thallium present in certain sulfide minerals is insignificant when compared with the thallium content of the silicate minerals of potassium, the new value may be regarded as applying to the total abundance of thallium in the upper lithosphere.

The purpose of this paper is to reconsider the newest abundance values presented for thallium and to compare the pair rubidium-thallium with some other pairs of elements closely associated in the upper lithosphere.

## MANNER OF OCCURRENCE OF THALLIUM IN THE UPPER LITHOSPHERE

Geochemically, thallium found in the upper lithosphere is characterized by its property of becoming enriched in late crystallates. In this respect it behaves like the heavy alkali metals, rubidium and cesium. The most important manner of occurrence of thallium in the upper lithosphere is its camouflage in silicate minerals rich in potassium. Potash feldspar and the micas, muscovite and biotite, evidently are the main seats of thallium in the silicic calcalkalic igneous rocks and particularly in their pegmatites. According to Ottemann (1941), thallium is also rather strongly enriched in the minerals of pneumatolytic tin-bearing rocks. Its occurrence in potash feldspar is explained by the fact that thallous thallium is able to form a feldspar structure, as the thermochemical calculations of Schiebold (1931) show. In potash feldspar, $Tl^+$ (radius 1.49 kX), like $Rb^+$ (radius 1.49 kX), substitutes for $K^+$ (radius 1.33 kX). No independent minerals of $Rb^+$ and $Tl^+$ are known to exist. In plagioclase feldspars, $Tl^{3+}$ (radius 1.05 kX) may become captured by $Ca^{2+}$ (radius 1.06 kX), as Ottemann (1941, p. 166) has suggested; in fact, Ottemann (1941, p. 162) reports as much as 10 g/ton thallium in plagioclase, whereas the maximum content found by Ahrens (1948*b*) is 2.9 g/ton thallium. In potash-rich silicate minerals, particularly in potash feldspar, thallium also accompanies cesium and lead. In like manner a geochemical alliance with lead characterizes the occurrence of thallium in sulfide minerals. Evidently, the most important site of thallium among sulfides is galena, in which it replaces lead. A similar replacement exists also in some plumbiferous sulfosalts, such as sartorite, $PbS \cdot As_2S_3$, dufrenoysite, $2PbS \cdot As_2S_3$, and jordanite, $4PbS \cdot As_2S_3$. The thallium minerals proper: lorandite, crookesite, vrbaite, and hutchinsonite, which are sulfides, selenides, and sulfosalts, are very rare and consequently do not possess much geochemical significance. Oftedal (1940) found the common sulfide minerals, with the exception of galena, devoid of thallium. The commonly reported occurrence of thallium in pyrite is probably erroneous (V. M. Goldschmidt, according to Oftedal, 1940). The average thallium content of primary magmatic sulfides, according to the analyses of Noddack and Noddack (1931), is 1 g/ton thallium. It is evident that the content of thallium in sulfide

minerals is too low to be of geochemical importance and to affect the abundance of thallium in the upper lithosphere.

## CONTENT OF THALLIUM IN IGNEOUS ROCKS

Ahrens (1948*b*) failed to determine the content of thallium in granites and other rocks relatively rich in potassium because their thallium content was below the limit of detection of the spectrochemical

TABLE 1

CONTENT OF THALLIUM IN IGNEOUS ROCKS

| Rock | Tl (g/ton) |
|---|---|
| Granites and liparites, average (Goldschmidt and Hörmann in Goldschmidt, 1937*b*) | 0.5–1 |
| Composite of 11 German gabbros (Preuss, 1941) | 0.3 |
| Composite of 14 German granites (Preuss, 1941) | 3 |
| Granite, Eibenstock, Harz, Germany (Ottemann, 1941) | 1.0 |
| Granite, Schellerhau, Harz, Germany (Ottemann, 1941) | 1.0 |
| Granite, Epprechtstein, Harz, Germany (Ottemann, 1941) | 2.0 |
| Granite, Brocken, Harz, Germany (Ottemann, 1941) | 2.1 |

method that he used. However, there is a number of direct previous determinations of the thallium content of igneous rocks, made by V. M. Goldschmidt and H. Hörmann (Goldschmidt, 1937*b*, p. 88), Preuss (1941, p. 19), and Ottemann (1941, p. 156). All these authors used spectrochemical methods of investigation; in order to increase the sensitivity, Preuss distilled the readily volatile elements from rock samples in a carbon-tube furnace directly into the carbon arc, and Ottemann heated the samples in a high-vacuum electric furnace, whereby sublimates were obtained for further studies. The results of their analyses are presented in table 1.

With due consideration to the scarcity of the analyses presented in table 1, the values show that the thallium content of subsilicic rocks is considerably lower than the content in silicic rocks. The high average reported by Preuss for the composite mixture of German granites may be attributed to the geochemical peculiarities of the Caledonian and Variscan granites analyzed. Some other trace elements are also known to be present in exceptional quantities in these granites. The average of Ottemann's analyses for the Harz granites is 1.5 g/ton thallium, a value which compares rather well with the average reported by Goldschmidt and Hörmann for granites and liparites. The values presented in table 1 clearly show that the new value given by Ahrens for the abundance of thallium in igneous rocks is much too high. But table 1 also shows that Goldschmidt's value is somewhat too low; the value 0.6 g/ton might be chosen to represent the true abundance of thallium in igneous rocks at least tentatively until more analyses become available.

The abundance of thallium and its neighbors in the Periodic System is given in table 2.

According to the rule of Oddo and Harkins, both mercury and lead should be more abundant than thallium. The abundance relationships of lead (abundance according to Goldschmidt, 1937*b*) and thallium certainly follow the rule, whereas in the case of mercury and thallium the relationship seems to be more obscure. The value 0.5 g/ton for mercury is based on determinations from sedimentary rocks (Goldschmidt, 1937*b*), whereas the lower value is the recent estimate of Saukov (1946); this value agrees well with the analyses of igneous rocks made by Stock and Cucuel (1934), Preuss (1941), and Saukov himself. The discrepancy may be explained by the assumption that mercury is geochemically

comparable with the atmophile elements which were largely lost during the early stages of the evolution of the earth. Therefore, the present terrestrial abundance of mercury is abnormally low. In like manner mercury is considerably more deficient in the meteorites than the rules for the cosmic abundance of its isotopes require (Suess, 1947, p. 604). As Goldschmidt (1937*b*, pp. 11, 84) points out, much of the mercury in the sediments and their derivatives probably has been produced by the degassing of the upper lithosphere, i.e., by volcanic emanations and hydrothermal solutions. If

TABLE 2

| Z | Element | Abundance in the Upper Lithosphere (g/ton) |
|---|---|---|
| 80 | Hg | 0.077–0.5 |
| 81 | Tl | 0.6 |
| 82 | Pb | 16 |

Ahrens' new value for thallium is considered, the deviation from the rule of Oddo and Harkins grows still more pronounced than with the lower values, and therefore it follows that a lower value is to be preferred. The rule of Oddo and Harkins, which is no selection principle but a true natural law, affords a welcome check of the abundance of the elements considered.

## THE Rb/Tl RATIO IN IGNEOUS ROCKS

According to Ahrens, the Rb/Tl ratio does not change during the crystallization of minerals. In order to study the Rb/Tl ratio in some igneous rocks, table 3 is presented; it is compiled from analyses available in literature. The first two rubidium values are from Goldschmidt, Berman, Hauptmann, and Peters (1933, p. 238); the third value is a redetermination by Goldschmidt, Bauer, and Witte (1934, p. 51). The thallium values are those listed in table 1, and the Rb/Tl ratio for the potassium-rich minerals of granite pegmatites is calculated from the mean $Rb_2O/Tl_2O$ ratio given by Ahrens (1948*b*, p. 580).

Table 3 shows that considerable variation in the Rb/Tl ratio occurs during the magmatic differentiation. Certainly, no conclusion on the abundance of thallium can be based on the assumption of the prevalence of a stable Rb/Tl ratio in igneous rocks and their minerals, and therefore the evidence of table 3 also contradicts the new high abundance value for thallium.

TABLE 3

CONTENT OF RUBIDIUM AND THALLIUM AND THE Rb/Tl RATIO IN SOME IGNEOUS ROCKS

| Rock | Rb (g/ton) | Tl (g/ton) | Rb/Tl (Weight Ratio) |
|---|---|---|---|
| Composite of 11 German gabbros | 18 | 0.3 | 60 |
| Composite of 14 German granites | 550 | 3 | 183 |
| Composite of 14 German granites | 830 | 3 | 277 |
| Granite pegmatites (potassium-rich minerals) | | | 95 |

The discrepancy between Ahrens' high value and the other, lower values is readily explained by the fact that the Rb/Tl ratio adopted by Ahrens as the universal ratio actually is valid only for the potassium-rich minerals of late crystallates, in particular of granite pegmatites, and not for the minerals of the igneous and quasi-igneous sequence as a whole. Moreover, the abundance of rubidium is not known with certainty, and therefore an abundance value for thallium calculated from the Rb/Tl ratio remains necessarily uncertain. It must be

emphasized that the abundance values given by Goldschmidt (1937*a*, *b*) for a number of elements are based on analyses of Caledonian and Variscan igneous rocks and consequently may be uncertain because there is much reason to believe that regional differences in the content of many trace elements exist in the upper lithosphere.

### THE COHERENT PAIR Rb-Tl COMPARED WITH SOME OTHER PAIRS OF ELEMENTS

Ahrens has emphasized that, with the possible exception of the pair Zr-Hf and, perhaps, some of the rare-earth metals, no other pairs of elements are known which are so closely associated as are "alkali-metal" thallium and rubidium in silicate minerals. With this statement Ahrens evidently refers to the concept of geochemical coherence: two or more elements are geochemically coherent if they constantly follow each other in nature, not only in igneous rocks but in the endogenic and exogenic cycle of terrestrial matter in general. A convenient measure of the geochemical coherence is the variation of the abundance ratio of the elements concerned in different materials. In the following the degree of coherence in some pairs of elements will be considered.

*Gallium-aluminum.*—As Goldschmidt and Peters (1931) showed, gallium always follows aluminum in nature. They found that in minerals the atomic Ga/Al ratio varies from about 1 : 10 to 1 : 120,000 —or by a factor higher than $10^4$. Consequently, the degree of coherence in the pair Ga-Al is low.

*Lanthanum and the lanthanides.*—This group of elements was formerly considered to be highly coherent, but recent research has proved that notable differences exist between the distribution of the various lanthanides, which may in certain cases become separated from each other (for a recent summary see Haberlandt, 1948).

*Columbium-tantalum.*—It is now known that columbium and tantalum actually may become quantitatively separated from each other in minerals (Rankama, 1944, 1948). Therefore, these metals do not form a fully coherent pair, even though they usually follow each other in nature.

*Zirconium-hafnium.*—This pair certainly is highly coherent: where zirconium does not occur, there hafnium is absent as well. Zirconium, consequently, is a very powerful protective element (*Schutzelement*) for hafnium. No hafnium is present in titanium and thorium minerals. It is an exclusive constituent of zirconium minerals. According to Hevesy (1925), the $HfO_2/ZrO_2$ ratio varies from 0.007 (in fava) to 0.5 (in thortveitite). The corresponding atomic Hf/Zr ratios are 0.004 and 0.3, respectively, and therefore the ratio varies by a factor of 75. The coherence is strong.

*Rubidium-thallium.*—According to Ahrens (1948*b*), the $Rb_2O/Tl_2O$ ratio varies from 11.3 (in muscovite No. 165) to 660 (in potash feldspar No. 16). The corresponding atomic Rb/Tl ratios are 27.3 and 1434.7, respectively, and the Rb/Tl ratio varies by a factor of about 53. It is evident that the coherence is strong.

A comparison of the data presented above shows that the coherence in the pairs Zr-Hf and Rb-Tl is higher than in the pairs Ga-Al and Cb-Ta, and probably also higher than in the lanthanide group, even though here the lack of analytical data so far does not allow a direct comparison. On the other hand, the coherence in the pairs Zr-Hf and Rb-Tl is of the same degree of magnitude. Even though the geochemistry of hafnium is known very inadequately, all

information available shows that hafnium never parts with zirconium in nature.

Of course, the values presented above allow a comparison of coherence only with reference to Zr-Hf– and Rb-Tl–bearing minerals of late crystallates, but it is probable that rubidium and thallium actually are a less coherent pair than zirconium and hafnium. The reason is that, all manners of occurrence and all valence states of thallium being considered, a complete separation of rubidium and thallium takes place in the thalliferous sulfide minerals. However, if only univalent thallium is considered, one may state, in material agreement with Ahrens, that the association of thallium and rubidium is, indeed, remarkably close, in the potash-rich silicate minerals of granite pegmatites and of silicic igneous rocks.

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# GEOLOGICAL NOTES

## TRIVOLI SANDSTONE OF WILLIAMSON COUNTY, ILLINOIS[1]

RAYMOND SIEVER
Illinois State Geological Survey, Urbana, Illinois

New information on the nature of the sandstones of the lower part of the McLeansboro group of the Pennsylvanian system in a part of southern Illinois has resulted from recent diamond drilling in northern Williamson County, in the southern Illinois coal field (fig. 1). A sandstone which crops out in the immediate vicinity of the bore holes is the uppermost sandstone encountered in the cores, and a definite correlation can be made from surface exposures to the subsurface section. Two sets of samples of this sandstone were collected, one from the drill holes, the other from the outcrop. Thin sections were cut and heavy mineral analyses made from these samples.

The sandstone sampled is the basal member of the cyclothem containing No. 8 coal. It lies unconformably on the uppermost shale member of the cyclothem next below. Its base is about 200 feet above the base of the McLeansboro group. Kosanke and Eddings (1946, p. 1212) have correlated coal No. 8 of southern Illinois with the Trivoli (No. 8) coal of western Illinois. Thus the basal sandstone member of the Williamson County cyclothem presumably can be correlated with the basal member of the Trivoli cyclothem, the Trivoli sandstone. Cooper (1945, p. 16) has correlated the Trivoli sandstone of western Illinois with the basal member of the Missouri series of the Mid-Continent section, lying just above a major floral and faunal break in the Pennsylvanian strata.

On the outcrop the sandstone is brown to reddish-brown, uniformly fine-grained, weathering into large slabs and flags, and uniform in appearance from top to bottom. The sandstone is prominently marked with oscillation ripples; the amplitude averages 2 cm., and the wave-length ranges from 8 to 12 cm. Cross-bedding of the aqueous type is common. The angle of the foreset beds is about 10°–15°. Characteristically the rock is case-hardened by hematite and limonite as much as 2 cm. from the surface. The basal conglomerate commonly present in the cores was not exposed in any outcrop.

In the core the sandstone is light gray, with a weathered brown zone at the top, commonly 2–5 feet thick. Shaly micaceous carbonaceous partings are not uncommon, and ripples are prominent. The top of the sandstone grades into the overlying silty shale member of the cyclothem through a transitional zone, 2–5 feet thick, characterized by thin shaly laminae in the sandstone, which increase in number and thickness upward. In all the drill holes the contact with the shale below is abrupt and uneven and, in many places, is inclined as much as 20°. The interval between the base of the sandstone and the top of the first coal bed below shows appreciable variation; the interval is commonly 50 feet, but in some places it decreases to 10 feet. It is probable that the Trivoli sandstone is unconformable on the shale below. In general, the sandstone is uniform throughout.

The sandstone ranges in thickness from 60 to 100 feet in northern Williamson County; in Franklin County, to the north, it reaches a maximum thickness of 157 feet. Approximately 6 miles north of the outcrop

[1] Published with the permission of the Chief, Illinois State Geological Survey. Manuscript received February 18, 1949.



FIG. 1
Illinois.

FIG. 1.—Index map, showing type area of Trivoli sandstone and location of Williamson County in Illinois.

area in Williamson County the sandstone grades laterally northward into sandy shale, and thence into shale. The sandstone is lost as an identifiable unit until it reappears some 60 miles to the north.

As shown in thin sections (pl. 1, *A*, *B*), the rock contains, on the average, 60 per cent detrital quartz grains. Next in order of abundance is the clay matrix, composed of sericite, chlorite, illite, and other clay minerals. The remainder of the rock consists of various types of feldspar, rock fragments, and chert. With a few exceptions, discussed below, chemical cements are less than 7 per cent. They include authigenic quartz, calcite, and siderite. Heavy mineral accessories are few and slightly varied, the most abundant being tourmaline and zircon.

The sandstone contains several types of detrital quartz. The most prominent is the normal igneous quartz which is free of strain shadows, has even edges, and contains small dusty inclusions. From 40 to 60 per cent of the detrital quartz is of igneous origin. Almost equally prominent are the metamorphic quartz grains which are lath-shaped, lensoid, and exhibit mild to pronounced strain shadows. Many of them are metamorphic quartzite fragments which show extreme cases of suturing of secondary quartz. Authigenic quartz as enlargement of detrital grains rarely exceeds 5 per cent of the total quartz. Second-cycle grains are scarce. The visual roundness averages about 0.25; metamorphic quartz is a little lower, and igneous quartz is a little higher, in roundness.

Feldspars, which constitute from 2 to 10 per cent of the rock, include sodic plagioclase, orthoclase, microcline, and a few grains of anorthoclase. The feldspars are slightly more rounded than the quartz and have a visual roundness of 0.5. Most of the feldspars are fresh, but some grains show extensive kaolinization.

Rock fragments, composing up to 15 per cent of the sandstone, are of many types of argillaceous sedimentary and low-rank metamorphic rocks and some chert. The most common fragments are phyllites. Many of the shale fragments are similar to Pennsylvanian shales lower in the section and thus may be of local origin.

The clay matrix constitutes about 20 per cent of the sandstone. It is exceedingly fine-grained, acting as a binder for the larger detrital grains. The most abundant mineral is sericite and, in order of decreasing abundance, chlorite, illite, and kaolinite. A few grains of biotite are present. The clay particles show no preferred orientation.

Muscovite is very prominent in the Trivoli sandstone. It occurs as long shreds and plates, many of them bent and broken, and shows conspicuous parallel orientation. Some samples contain as much as 5 per cent of muscovite.

Two carbonate minerals—calcite and siderite—occur as cement. The carbonate cement composes about 5 per cent of the rock, except for a few samples in which it is as much as 20 per cent. The carbonate occurs as irregular patchy areas and replaces quartz and feldspar. In the few samples with high carbonate content the rock has a relatively clean appearance; the clay matrix is at a minimum, and calcite is the dominant cement. This calcite has extensively replaced quartz and feldspar and is itself being replaced by authigenic quartz. Presumably the clean-washed permeable sand was more favorable to carbonate infiltration and precipitation. Sharply euhedral secondary quartz replaces carbonate. Siderite occurs as small irregular clusters of rhombohedrons

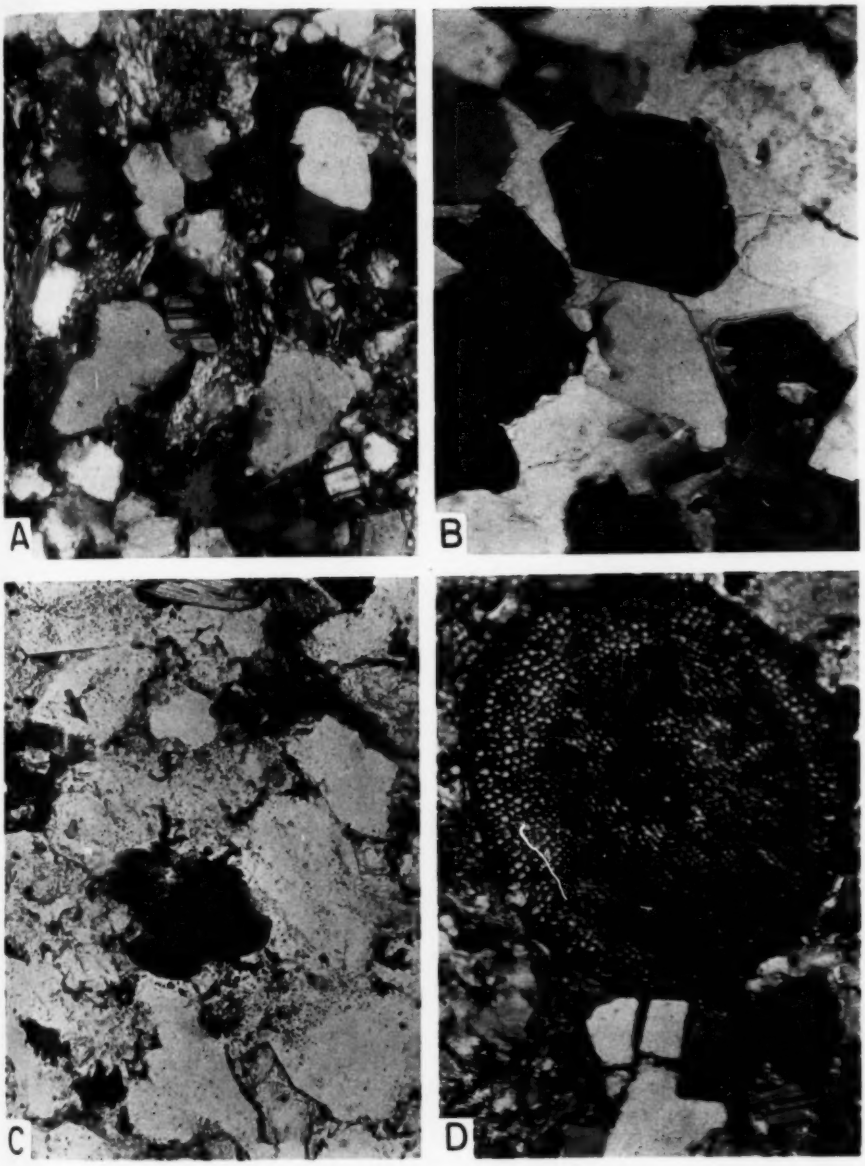
PLATE 1

*A*, Typical appearance of Trivoli graywacke. Quartz, plagioclase feldspar, and clay matrix are readily distinguishable. Crossed nicols. Magnification 100×.

*B*, "Washed" graywacke with calcite cement. Euhedral secondary enlargement of quartz replacing calcite cement. Crossed nicols. Magnification 100×.

*C*, Glauconite (*arrow*) in the Trivoli sandstone. Plane-polarized light. Magnification 100×.

*D*, Cross section of fusainized stem in the Trivoli conglomerate. Crossed nicols. Magnification 100×.




Thin sections of Trivoli sandstone

and as extensive replacement of calcite. The siderite is about 30 per cent of the carbonate cement. It is extensively altered in the outcrop to hematite and limonite.

A very small amount of glauconite (pl. 1, *C*) is present. As it shows no evidence of being worn in transport, it is considered to be authigenic. It occurs in samples with a greater amount of biotite than customary and, as suggested by Galliher (1935, pp. 1351–1366), may be the product of marine alteration of the biotite.

The basal conglomerate of the sandstone, found in about 65 per cent of the drill holes, is generally 2–3 feet thick, but rarely attains a thickness of 6 feet. The matrix and fine detrital material are much the same as in the nonconglomeratic sandstone, but the conglomerate has slightly more carbonate cement, as much as 15 per cent. The pebbles are irregular in shape, commonly flattened, with the long dimension ranging from $\frac{1}{4}$ inch to 2 inches. Almost all the pebbles are of clay ironstone concretions, such as are commonly found in shales above a coal bed in Illinois. Other pebbles are of a sandstone tightly cemented with hematite and limonite, an occasional limestone pebble, and rarely a coal pebble.

Noteworthy are the plant remains found in the conglomerates (pl. 1, *D*). They are not impressions but are large fragments of stems and twigs whose abraded edges show evidence of transport. Invariably, the material is fusain or semi-fusain. The intracellular pore spaces of the fusain are open, in contrast to the fusain in Illinois coal beds, which is often mineralized with pyrite.

The mineralogy of the Trivoli sandstone is in sharp contrast to the mineralogy of sandstones lower in the Pennsylvanian section in Williamson County, although the Trivoli sandstone of Williamson County is similar to the sandstones between the Isabel sandstone at the base of the Carbondale group and the Trivoli sandstone of western Illinois. The lower sandstones in Williamson County average 80 per cent detrital quartz and 20 per cent chemical cement; in general, they are washed clean. They are typical representatives of the clean quartz sands of the orthoquartzite series. Thus the mineralogy of the Trivoli sandstone may be a reflection of the uplift and renewed erosion of the source area. The conglomerates of the lower Pennsylvanian in southern Illinois are very different from the Trivoli conglomerate. The former consist of thick massive beds containing well-worn quartz pebbles transported from a distant source; the latter is a thin basal conglomerate of slightly worn material of local derivation.

According to the cyclothem hypothesis of Pennsylvanian sedimentation, most of the basal sandstones of cyclothems have been considered to be continental deposits (Weller, 1930, p. 119). The few exceptions pointed out have been interpreted as "the shoreward facies of marine deposits contemporaneous with the aggradation of the alluvial plain" (Weller, 1930, p. 120). A major objection to the marine origin of most of the sandstones is the occurrence of autochthonous coal beds in them. No coal beds have been observed in the Trivoli sandstone of Williamson County. Petrographic and field evidence suggests that this sandstone in this particular area is more probably marine than nonmarine. The occurrence of wave ripple-marks, the uniformity of lithology, the thin basal conglomerate, and the glauconite are indications of a marine environment. This sandstone appears to constitute a marine facies of a dominantly continental deposit.

The sandstone is in the category of a low-rank graywacke or a subgraywacke. It implies contemporaneous moderate deformation in the source area of the sediments. The major source area has been considered to be "Appalachia," with minor amounts of detritus coming from Llanoria and the Wisconsin highlands. Investigations are in progress to establish more definitely the sources of sediment, the influence of tectonic movements of the source areas on sedimentation, and the amounts contributed by each source area to the sediments of the slowly downwarping Eastern Interior basin in Pennsylvanian time.

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## A PROPOSED METHOD FOR THE MEASUREMENT OF GEOLOGIC TEMPERATURES[1]

I. I. FRIEDMAN

University of Chicago

The establishment of an absolute geologic thermometer is of utmost importance in modern petrology. At present it is possible only in rare cases to determine with any degree of certainty the temperature at which a rock has been formed. If a means of temperature determination accurate to 100° C. and usable from 100° to 1000° C. were available, many of the problems of metamorphic and igneous petrology could be solved. For example, the question of the solution or replacement origin of pegmatite bodies might be answered if their temperature of emplacement, as well as the temperature of emplacement of the enclosing rocks, could be determined. One of the biggest problems of petrology, namely, the origin of granites, also hinges upon questions of temperature of formation. At present most methods of temperature determination are based upon the temperature stability of various minerals or mineral assemblages. These methods have the disadvantages of giving only a relative measure at best, and one the value of which is a function of composition as well as of temperature and pressure.

The following method has the advantage of being a first-order function of the temperature and pressure and depends on bulk rock composition in but a minor way.

The method, simply stated, is to measure the distribution ratio of a suitable minor element between two minerals in equilibrium in a rock. The distribution ratio should be determined solely by the temperature and pressure at which the distribution was established.

The minor element should be one that is present in large enough quantity for accurate and simple determination. This element should not be one that substitutes easily in one of the mineral lattices and not in the other, although this qualification may prove to be unimportant. One important requirement is that the minor element be present in such small amounts that the laws of the perfect solution can be applied.

H. C. Urey[2] has suggested the use of an ion which will substitute for an ion present in both minerals, i.e., germanium substituting for silicon in quartz and feldspar. Such a system is more likely to behave like a perfect solution than is the case in which large lattice distortions result upon substitution.

Because only small amounts of each phase can be obtained in a pure condition because of the fineness of the grain of many of the rocks in question and because the element to be determined is present in a concentration of 0.01–100 parts per million, a very accurate and sensitive analytical tool is necessary. The radiochemical method (Brown and Goldberg, 1949, p. 347) fits the above requirement. The work can be divided into several parts as follows:

1. An element must be found that exists

[1] Manuscript received August 8, 1949.

[2] Personal communication.

in natural rocks in the proper concentration and whose distribution coefficient is not too far removed from unity. This element must be one whose concentration can be conveniently determined with sufficient accuracy. In addition, the slope of the temperature versus distribution coefficient curve must be sufficiently large that the proper accuracy can be obtained.

2. The temperature coefficient of the distribution ratio between two minerals—say, quartz and potash feldspar—must be determined. This may be done in two ways: (*a*) by the simultaneous synthesis of the two minerals in a closed system at a known temperature and pressure and (*b*) by equilibrating a natural rock at different temperature and pressure in the presence of water vapor.

3. Field investigations—a series of rocks from whose occurrence a relative temperature scale can be derived—must be investigated to see whether equilibrium relationships occur as far as the distribution coefficient is concerned.

4. The influence of solid solution and order-disorder changes in one host lattice on the distribution coefficient must also be investigated.

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# REVIEWS

*Cretaceous and Tertiary Subsurface Geology: The Stratigraphy, Paleontology, and Sedimentology of Three Deep Test Wells on the Eastern Shore of Maryland.* (State of Maryland, Department of Geology, Mines, and Water Resources, Bull. 2 [1948].) Pp. 456; figs. 30; pls. 39. $2.00.

This bulletin is a comprehensive report on three deep tests drilled during the period 1943–1946 by the Ohio Oil Company, the Socony-Vacuum Oil Company, and the Standard Oil Company of New Jersey. Cuttings, cores, and electric logs were donated to the Department of Geology, Mines, and Water Resources, under the directorship of J. T. Singewald, Jr. Several collaborators conducted an exhaustive study of the data furnished by the wells, especially a complete core from a depth of 1,000 feet to bottom (5,568 feet), in the Ohio Oil Company—L. G. Hammond No. 1 well in Wicomico County, Maryland.

The first one hundred and twelve pages of the volume are a report on the subsurface Cretaceous and Tertiary geology by J. L. Anderson, based largely upon the cored well and tied in with other data. Detailed heavy-mineral and size analyses were run on core samples, affording a wealth of data in the form of histograms, statistical parameters of size, and graphs of the relations among the parameters. These data, in conjunction with the paleontological evidence, were used to reconstruct the environment of deposition of the sediments and to correlate the section among the three wells. Ditch samples from the Maryland Esso well were logged by R. M. Overbeck, and an appendix at the end of the volume presents the log data.

The second part of the volume describes the paleontology of the section penetrated by the cored well. Experts in several fields were engaged in the study, including Julia A. Gardner, Ann Dorsey (Mrs. F. W. Clapp), L. W. Stephenson, H. E. Vokes, K. E. Lohman, F. M. Swain, and J. A. Cushman, under arrangements made with Dr. J. B. Reeside, Jr., of the United States Geological Survey. The papers on paleontology include studies of mollusks, diatoms, ostracods, and Foraminifera. Many new species were found, which are described and illustrated in the thirty-nine accompanying plates. These papers provide an important reference on the paleontology of the East Coast Tertiary and Cretaceous, which will prove valuable also to workers along the Gulf Coast.

The data from the three wells, tied in with outcrop and other well information, show that the section thickens markedly from west to east. In Kent County, just east of Chesapeake Bay, the Tertiary is about 100 feet thick. In Worcester County, along the coast some 80 miles southeast, the Tertiary is 2,000 feet thick. The Cretaceous is 1,200 feet thick in Kent County, and nearly 6,000 feet thick in Worcester County. Between the Cretaceous and the Basement Complex east of Chesapeake Bay is a buried occurrence of Triassic rocks, consisting mainly of reddish-brown and apple-green shales, sandy shales, and arkosic sandstones. The correlation is based on lithologic similarity to the known Triassic of Marlyand, inasmuch as no fossils were found. The Triassic is 585 feet thick in Worcester County, where it is more shaly than in the Ohio Oil Company well farther southwest.

The conditions of sedimentation were reconstructed mainly from data afforded by the cored well. It was concluded that the beds in this well were deposited in a continental environment which changed gradually to deltaic and finally to marine. Marine conditions set in at the base of the Matawan formation (top of Upper Cretaceous) and continued uninterruptedly into the Tertiary. Progressive overlap of the beds toward the west is more pronounced in the Lower Cretaceous than in the upper part of the well section. Some brackish-water conditions in the Upper Cretaceous were noted in the easternmost of the three deep wells.

The petroleum potentialities of eastern Maryland were assessed from the well data, and it was concluded that, although the section is suitable for the accumulation of petroleum, it is not very favorable for petroleum generation.

If oil occurs in the Cretaceous, it is believed that it will be found east of the coast line, in more marine parts of the section. The Tertiary, although marine, has only shallow cover and seems unsuitable for oil occurrence. No evidence of petroleum was found in any of the sands in the three wells.

It is apparent from the detailed studies made upon the material and from the critical analysis of the data afforded that this volume provides a definitive treatment of the subsurface Tertiary and Cretaceous of the central part of the Atlantic Coast. Its value lies not alone in the insight that it affords into the geological history of an important area but also in the comparative data which it affords for regional studies of the Cretaceous in the entire coastal overlap area. The volume will undoubtedly become a standard of reference; and the Department of Geology, Mines, and Water Resources, together with the co-operating companies, is to be complimented on making this information available.

W. C. KRUMBEIN

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"Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit." By HANS POSER. (*Erdkunde*, Bd. 2, Lfg. 1–3.) Pp. 53–68. Bonn, 1948.

This paper is a co-ordinated condensation of two articles published in *Die Naturwissenschaften* (Bd. 34, pp. 10–18, 232–238, 262–267, 1947). It deals with the time of maximum of the last European glaciation, the Würm-Weichsel (Brandenburg stage). Perennially frozen ground or pergelisol[1] in strict sense and inferred temperature conditions are discussed in the first part. In central and western Europe there is evidence of glacial-age pergelisol in loess and clay wedges, involutions (*Würgeböden*, *Taschenböden*), and climatically induced asymmetric valleys. Loess and clay wedges are superficial features which form nets with large meshes. They represent fillings of ancient ice-wedge nets which develop at very low temperatures. Involutions are excessively disturbed, squeezed, and disrupted structures of originally horizontal beds and are forming today in the pergelisol tundras of Siberia but not in eastern or central Europe. They are caused by pressure exerted through freezing from below and above of the ground thawed in summer, the mollisol. Climatically induced asymmetric valleys, usually with steep west- and south-facing slopes, occur only in the headwater regions (of central Europe). As known from Siberia, such valleys form because slopes facing west and south thaw out sooner and deeper than do slopes facing in other directions and therefore erode more easily.

Data on these three indicators of glacial-age pergelisol in western and central Europe are compiled on a map (fig. 1). On the whole, their northern limit of distribution coincides with the southern boundary of the Weichsel (Brandenburg stage) glaciation. Involutions which occur in places in the region of young drift date from the ice advance. The equatorial border of the pergelisol was mainly dependent on the depth of the snow cover and on the winter temperature and may imply a mean annual temperature of less than $-2°$ C. (28°.4 F.; also see Stephen Taber in *Geol. Soc. America Bull.* 54, p. 1506, 1943). It passed from Brittany to the Alps and through northern Yugoslavia and southern Rumania.

The Würm-age timber line, which, like its modern successor, may have coincided with the July-isotherm of 10° C. (50° F.), is taken from the work of Franz Firbas. This line ran somewhat south of the pergelisol limit in France; west, south, and east of the Alps to Czechoslovakia; and then southeastward on the south side of the Carpathians. These boundaries divide central Europe into four main ancient climatic regions: (1) the pergelisol tundra extending from Land's End into Russia; (2) the continental pergelisol forest east of the Alps; (3) the maritime pergelisol-free tundra which forms a narrow strip in France; and (4) the maritime pergelisol-free forest west of the Alps (fig. 6).

At the intersection of the pergelisol and timber lines in northwesternmost Yugoslavia,

[1] The term "pergelisol" was proposed, together with "involutions," "plications," "mollisol," etc., by Kirk Bryan in the *Am. Jour. Sci.*, vol. 244, pp. 622–642, 1946, for perennially frozen ground and to replace "tjäle" and "permafrost." However, as Carl Troll stresses in *Erdkunde*, Bd. 2, p. 6, 1948, intensive frost action is by no means limited to perennially frozen ground but is also produced by winter, short periodic, and nightly freezing of the ground. Troll therefore proposes the additional terms "annual gelisol" and "diurnal gelisol." He states that active annual gelisol is present in the Scandinavian mountains, active diurnal gelisol in many mountains of low and middle latitudes, including those in New England (Ernst Antevs, *Alpine Zone of Mt. Washington Range* [Auburn, Maine: Merrill & Webber Co., 1932], pp. 34, 35, 40, 57, 61).

where the July temperature may have been 10° C. (50° F.) and the annual mean −2° C. (28°.4 F.), the temperature is estimated to have been 8°–9° C. (14°.4–16°.2 F.) lower than at present in July, and 12° C. (21°.6 F.) lower in January. In the strip between the two boundaries in France the temperature may have been about 7° C. (12°.6 F.) lower in July, and 0°–3° C. (0°–5°.4 F.) lower in January.

The second part of the paper deals, first, with the depth of the summer thaw (thickness of the mollisol) during the last glacial maximum and inferred summer climate. The depth of the seasonal thaw is recorded by involutions, plications, or folds (*Stiche*) and, with reservations, by loess and clay wedges. All three features give minimum values. To eliminate topographic influences, only records from relatively level land less than 300 meters (984 feet) in elevation are used. The data are plotted on a map (fig. 2), on which three regions become apparent. The forested Hungarian Basin thawed to a depth of only 1.5–3 meters (4.9–9.8 feet), probably because the vegetation exerted a shielding effect, whereas the tundra from Denmark to the Seine River thawed to 1–3 meters (3.3–9.8 feet), suggesting a long season with positive, even if low, temperature. Central Germany had a thaw depth of only 1 meter, apparently because of strong cooling by the Nordic and Alpine ices.

These temperature conditions and their distribution give indications about the summer pressure, wind, and precipitation. The high atmospheric pressures over the ices may frequently have been connected by a bridge, while low pressures may have prevailed over the Hungarian Basin and northwestern continental Europe (fig. 3). In the region east of the Oder River and the Alps the prevailing winds came from the northeast; in western and central Europe from the southwest. Cyclonic storms must frequently have progressed from the Mediterranean over Hungary and from the Atlantic over France, the Low Countries, and Germany. Precipitation was probably heaviest at the ice border and in central Germany, though this latter region by some students is assumed to have had a cold-dry climate.

Whereas the depth of thaw bears on the ancient summer climate, disruption of already frozen ground by contraction during very low temperatures gives hints on the winter climate. The intensity of the frost tearing can be judged from the depth and relative abundance of frost cracks and clay wedges, these latter requiring lower temperatures. In northwestern continental Europe are found only narrow and, at most, 3-meter (9.8-feet) deep frost fissures, whereas in the interior there are some cracks but predominantly clay wedges, which in central Germany are mostly over 5 meters (16.4 feet) deep (fig. 4). It is concluded that the mean January temperature in the Hungarian Basin was somewhat lower, in central Germany decidedly lower, and in the maritime northwest part of the continent distinctly higher than that at the intersection of the pergelisol and forest limits in northwestern Yugoslavia, where it is calculated to have been about −14° C. (6°.8 F.). From these temperature conditions it is, in turn, concluded that the atmospheric pressure was high over most of Europe and that an offshoot of low pressure extended in over northern France and contiguous regions (fig. 5). Southwest winds may have prevailed over western Europe, east and northeast winds in the region east of the Elbe River and the Alps. Except in the west, the pressure conditions may have been very stable, cyclonic storms few, and precipitation small.

The main results of this important study of the ground-frost and climate during the Weichsel-Würm glacial maximum are summed up in a map (fig. 6), purporting to show the then climatic provinces of middle and western Europe, and in an accompanying brief description of the provinces.

ERNST ANTEVS

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*Soil Mechanics in Engineering Practice.* By KARL TERZAGHI and RALPH B. PECK. New York: John Wiley & Sons, Inc., 1948. Pp. xviii+566. $5.50.

The book on *Soil Mechanics in Engineering Practice* by Terzaghi and Peck should prove of general interest to geologists. The science of soil mechanics has developed very rapidly in the last ten years. Since the publication of Terzaghi's work on *Erdbaumechanik* more than twenty years ago, he has been the leader in this field. Thus this book is particularly timely. Engineers in recent years have become increasingly conscious of the need for precise information upon the behavior of soft or loosely consolidated material when it is subjected to loads or stresses. Engineers cannot always pick and choose the places where they are asked to build structures and do their work, so they must do the best they can under the conditions with which they must work.

This book presents a very thorough analysis of present concepts of the principles of soil mechanics and indicates in considerable detail how these principles can be applied effectively to engineering problems. It also includes sections on methods of procuring and testing samples for analysis.

Soil mechanics is useful to the geologist as well as to the engineer. First, it helps in his consulting work on engineering problems, because all the practical applications of soil mechanics to engineering have geologic aspects, of which the geologist should be aware. Such engineering problems as adequacy of the earth to support structures, including both buildings and dams; settlement of the ground after structures are built; stability of slopes of cuts and embankments; earth pressure on retaining walls; strength of earth materials in dams; highway and airport construction; and tunnels in soft ground all concern a geologist at one time or another.

Second, soil mechanics helps the geologist in understanding the mechanics of deformation of rocks, because the basic principles of mechanics apply equally well to hard rocks as to soft rocks. The principles of soil mechanics are particularly useful to the geologist in interpreting the folding and faulting of sequences of interbedded soft and hard rocks and also in studying landslides and hill creep.

Third, compaction of sediments is closely related to the problem of what the soil mechanics engineer calls "consolidation." The engineers have made more progress on compaction than the geologists have, and the latter can profitably read the excellent discussion of this problem in Terzaghi and Peck's book.

The fundamental basis of soil mechanics is the deformation of loosely consolidated earth materials when subjected to stress. Basically, it involves (1) the movement of the water contained in the sediment and (2) the movement of both the water and the solid constituents.

The problem is approached by a study of the hydrodynamic and physical principles involved and by measurements of sediments under field and laboratory conditions. Though the soil mechanics engineer attempts to attack the problem dynamically, his approach thus far has been to a considerable extent one of statics and of comparison with previous experience. The geologist therefore, because he is constantly being compelled to consider dynamics—that is, the changes that take place with time—and because of his knowledge of the habits of sedimentary materials, is in a position to assist the soil mechanics engineer. Terzaghi and Peck point out clearly this relationship between geology and soil mechanics.

The fundamental geologic principles influencing the behavior of water in sediments or in loosely consolidated earth, as this reviewer sees the problem, are: (1) grain-size distribution and shape of particles; (2) nature of mineral constituents, particularly the clays; (3) chemical composition of the water in the sediments, especially with respect to possible base exchange reaction; (4) the stress that is about to be or has been applied to the sediments; and (5) the influence of time, particularly as to past and future rates of change of stress. Of course, it will take a long time before these fundamental principles are understood thoroughly; yet their solution should be an objective of both geology and engineering. Meanwhile, Terzaghi and Peck have given us many fundamental and empirical relationships that can be applied very satisfactorily to engineering and geological practice.

PARKER D. TRASK

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*Physical Geology*. By CHESTER R. LONGWELL, ADOLPH KNOPF, and RICHARD F. FLINT. 3d ed. New York: John Wiley & Sons, 1948. Pp. 602; figs. 365. $5.00.

This is the third edition of a well-known and widely used elementary geology text which appeared in two previous editions (1932 and 1939) under the title *Textbook of Geology*, Part I: *Physical Geology*, by the same authors. The present revision has not involved radical changes in the organization or subject matter; its most striking changes are physical. Instead of the dull maroon or dark green of the previous editions, the present book is bound in the bright crimson and black of the authors' *Outlines of Physical Geology*. The paper on which the book is printed is of far better quality, and on it the photographs reproduce exceedingly well and are also larger and clearer than those in the previous editions. Appropriate new photographs have been added, and there are more of the excellent aerial photographs which distinguished the former editions. The "bleed" arrangement of many of the photographs gives the book a fresh, modern look. Most of the diagrams have been redrawn to give greater clarity or are wholly new.

A notable departure of this edition is the set of 250 original kodachrome slides which were made to accompany it. For years slides have been used to illustrate features and processes in elementary geology; they have usually been dull, black and white (or more often, gray) "lantern slides." In many instances professors, even those well along in years, showed their students the same slides that they had looked at as students in elementary courses—slides often featuring a whiskered geologist in a derby hat or ladies in the costumes of 1900 or thereabouts. The work of conversion of departmental collections to kodachrome slides lagged far behind the popularity of this type of photography, for it is an arduous task. Such a collection as that offered to illustrate this book (price $85.00 with carrying case and manual keying them to particular topics in *Physical Geology*) should be a boon to teachers of elementary geology courses, particularly those in small departments. In a larger department they can well form the nucleus around which a larger collection can be assembled. The slides, available only in the complete set, are made of film described as "unbreakable" and bound into plastic mounts. They can be used in an ordinary kodachrome projector. The photographs were taken by Orlo Childs of the geology staff at the University of Wyoming, who visited classical localities in all sections of the country. An excellent idea that he followed was to take the pictures, wherever possible, in well-known vacation spots, thus drawing on past and future experiences of the average student.

To the book a new chapter has been added, chapter ii, "The Method and Scope of Geologic Studies." In line with modern trends in science education, this chapter contains, first, a discussion (perhaps too brief) of the scientific method, leading to the idea that geology advances through reasoning based on accurate observation of features and processes, with the goal of reading a complete history of each geologic feature and ultimately of earth itself. With a bow to methodology, the writers state that conclusions are significant only as the methods by which they have been reached are fully understood. They believe that, for the elementary student at least, understanding the basic concepts of geology is more important than memorizing a large number of details. (And, incidentally, series of tests on college students have shown that understanding of concepts is far more lasting than knowledge of specific facts, which are fast forgotten.) Following the policy of the other editions, and with this in mind, the writers have placed more of the purely factual background material or "tools" (minerals, rocks, topographic map explanation, and the time scale) in appendixes, with the hope that this material will become the nucleus of laboratory work, inserted where it is needed in the course for understanding the development unfolded in the text.

In a masterly stroke, Dr. Longwell, responsible for chapter ii, illustrates the scientific method by discussing, almost as the student begins the book, a major problem still partly in the "twilight zone." This problem is the relation between continental masses and the deep-sea floors—in other words, an introductory discussion of the broad causes of diastrophism. Not only does this set the stage for the student by showing him an essential problem of geology and of earth's future, but it gives him a large, oriented picture of our planet's processes and problems, bold in outline, on which he can, from time to time, paint in, in their proper places, the concepts and ideas which he learns throughout the book. When he finishes the course, he thus should have made for himself a large, meaningful, integrated picture rather than a series of those perhaps interesting, but unrelated, small pictures which form the collection of many students when they finish textbooks or courses in elementary geology. Even more important from the standpoint of student interest, without which the instructor is powerless to teach effectively, is the fact that a big, important, "dignified" problem has been presented. In many texts and courses on geologic processes the student must spend considerable time, particularly in the early parts of the course, on topics like "work of the wind," "weathering," "work of running water," which, in many cases, fail to catch his interest and respect—because he cannot yet see the magnificent whole into which they will fit. He is likely to contrast it unfavorably with the courses he is taking concurrently —physics, "glamorized" by the introduction of atomic energy; astronomy, where he deals almost immediately with vast distances and majestic vistas of space; chemistry, where he studies the ultimate nature of matter. Dr. Longwell's presentation is skilfully made; the student is given information that he can understand at this point and an incentive for learning the de-

tailed evidence and following the detailed reasoning of the body of the book.

LOU WILLIAMS PAGE

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*The Religion of the Modern Scientist (Neo-materialism)*. By S. W. TROMP. Leiden, Netherlands: A. W. Sÿthoff's Uitgeversmaatschappy N.V., 1947. Pp. xxiv+480; figs. 113; 1 table.

The author of this broadly inclusive volume is a geologist, still relatively young, who has spent several years in petroleum exploration in the Dutch East Indies, Egypt, and Turkey and is at present the professor of geology in the Fouad I University in Cairo. His work will therefore be of great interest to all geologists, especially so because more than half the book is devoted to an exposition of geologic principles and an account of earth history. As a matter of fact, the title of the book is likely to be misleading; not that geologists should shy away from it if it were really what one would naturally expect it to be. Actually, there is little in it that can be called "religion," although there is quite a bit of philosophy and much that might serve as a good foundation for an intelligent code of personal behavior.

It is Dr. Tromp's thesis that "the law of action and reaction which determines the absolute causality in the non-living world must be applied also on the living world, being ruled by the same neo-materialistic laws." To demonstrate the correctness of this idea, he presents a digest of all knowledge, running the gamut from the "wonders of the crystalline world" through the minerals and rocks, the structures and history of the earth, the evolution of plants and animals, to mental processes, telepathy, and psychokinesis. With such an ambitious program it is almost inevitable that an occasional erroneous statement crops out here and there. For example, in discussing "experiments with tissue cultures in vitro" on page 199, he describes "vitro" as "a special feeding-fluid."

It is presumably Dr. Tromp's interest in "absolute causality" that impels him to search for "laws" in all aspects of cosmic evolution. What he means by "law" is perhaps illustrated by his "law of varying strike line of oscillations." This is one of his "main laws of mountain building." He defines it as follows: "The strike line of the oscillatory movements in the same region is often subject to variations during the different geological periods." If that is a "law," it denotes a concept of "law" that makes it possible to accept many of his statements pertaining to the deeper problems of human life, such as this one on page 241: "There is no indication that the thinking process of man is ruled by laws fundamentally different from those ruling the crystalline inorganic world."

It is, furthermore, quite logical for him to conclude that "certain symptoms in the hand caused by the distribution of nerve centers might indicate certain phenomena which are going to take place during the future life of that person" and that there may well be something in astrology and the prophecies of mediums. Pursuing that line of thought, he devotes several very interesting pages to a consideration of dowsers and divining rods. Even though one may not agree with his conclusions—and most certainly I do not—the sympathetic analysis that he gives is quite informative and well worth reading.

Even though I cannot agree with Dr. Tromp that the fatalistic determinism of his "neo-materialism" is the religion of the modern scientist, I am glad to recommend most highly his synthesis of the geological sciences. His summary of geologic life-development and of human history from Pleistocene time into the new age of atomic energy, as well as his critique of geologic principles, is cogent and refreshing. It is not seriously marred by the numerous typographical errors and the non-American idiom that frequently diverts the attention of the reader from ideas to words.

KIRTLEY F. MATHER

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*Mining Geology*. By H. E. MCKINSTRY, with sections by STANLEY A. TYLER, E. N. PENNEBAKER, and KENYON E. RICHARD. New York: Prentice-Hall Book Co., Inc., 1948. Pp. 602, plus 57 pp. of appendixes and glossary containing tables of rock classifications, chemical elements, natural functions, etc., and familiar mining and geologic terms; figs. 149; pls. 2; tables 13. $10.00.

This book is a constructive contribution to the progress and development of the application of the science of geology to mining. Here is an orderly, rational, and lucidly written exposition of the ABC's of where and how to look for factual data and how to collect, present, and

evaluate them when examining an ore deposit. Moreover, the book explains how these data may be used for guidance in reaching the economic decision as to whether it is an ore deposit from which one or more metals or minerals can be recovered at a profit or just another mineral deposit of no immediate economic importance.

The author divides his book into four parts: (I) "Assembling Geologic Data"; (II) "Geological Principles of Ore-Search and Ore Apprisal"; (III) "Application in Specific Phases of Mining"; and (IV) "Technological Characteristics of Ores." Individual chapter headings include such varied subjects as "Geologic Mapping," "Sampling Ore and Calculating Tonnage," "Drilling," "Guides to Ore-Targets and Loci," "Valuing Mining Properties," "Writing and Reading Reports," "Geological Work at an Operating Mine," "Marketable Forms of Metals and Ores," and a "Glossary of Mining and Geological Terms." The book is well documented with references. As one could expect at a price of $10.00, the book is well illustrated, printed, and bound.

This treatise gives an excellent summary of the responsibilities and duties of the mining geologist. It differs from the many treatises on "mineral deposits," "economic geology," and other textbooks, in that it is devoted wholly to geology. The author does not hesitate to orient and focus his whole effort on the industrial application of the embryo science. Not so much as a paragraph is devoted to questions of origin or other theories whose practical application cannot be foreseen.

The book is primarily concerned with the everyday tasks of the mining geologist—mine-mapping, logging and interpreting diamond drill cores, and analyzing and interpreting geology primarily for the purpose of guiding mine prospecting and exploration. The emphasis is on the search for ore and ore-finding more than on mine-examination work, although, to be sure, one involved strictly in mine examinations should benefit considerably by the geologic approach presented here, for the author approaches this question from a geologist's point of view rather than from a mining engineer's viewpoint. There is already available a long series of excellent books by mining engineers that treat various phases of mine examination, but only Forrester's recent book, *Principles of Field and Mining Geology*, is also written from a geologist's point of view.

While engineering technology (production economics) is unquestionably a vital factor in mine examination, the basic problem may be geologic. It benefits the engineer little to discuss at length in this report the relative economic merits of square-set stoping versus top slicing if his sampling and appraisal of ore reserves is based on wrong geologic assumptions. The selection of the stoping method will, moreover, probably be dictated by geologic factors, such as character and strength of the rock overlying the ore, presence or absence of slips and sheeting, etc. Many concentrating problems are also influenced, if not controlled, by details of the ore occurrence, as the author points out. McKinstry's geologic presentation of the problem is timely.

While the experienced mining geologist who has read widely in his profession may find few new ideas in the book, I am sure he will benefit by the orderly presentation and logical development of the author's thesis. We need more books like this, which show the young geologist how to apply the theory that he has learned in school and how to develop the field application of his profession. I recommend the book to all graduate students in economic geology not only for the sake of its thesis but also as a splendid example of simple, well-organized, lucid, and interesting writing on a technical subject.

OLAF N. ROVE[1]

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*The Iron Resources of California.* (*California Division of Mines Bull. 129.*) 1948. Pp. 304.

This book is a valuable résumé of the work done on the iron-ore deposits of California by the United States Geological Survey and the United States Bureau of Mines during the recent world war. Thirteen of the more than one hundred deposits of iron ores known in California are described in detail; an article on the titaniferous magnetites of the western San Gabriel Mountains in Los Angeles County is included, as is also a summary of the present iron-ore situation in California and a short section describing the results of the diamond drilling conducted by the Bureau of Mines on five of the properties. The thirteen deposits described were selected because it was thought that they were the largest and had the greatest possibilities for commercial development. The deposits are the Lava Bed district, the Silver Lake district, Old

[1] Published with the permission of the Director, U.S. Geological Survey.

Dad Mountain, Cave Canyon district, the Vulcan deposits, the Iron Hat district, Ship Mountains, and the Kingston Range, all in San Bernardino County; the Minarets deposit along the crest of the Sierra Nevada Range in Madera County; the Shasta and California deposit and the Hirz Mountain deposit in Shasta County; and the deposits at Lake Hawley and Spencer Lakes in Sierra County. The five deposits drilled by the Bureau of Mines are Eagle Mountains, Silver Lake, Lava Bed, Shasta and California, and Minarets.

The total iron-ore reserves of California are estimated at between 100,000,000 and 150,000,000 long tons, of which about half is not now commercial, either because of the long distance to a market or because of the small size of the deposits. The Eagle Mountains area contains about a third of the total reserves and is the only deposit in California capable of supporting a modern blast furnace for a period of ten years or more. Reserves of all the other deposits examined, except that at Kingston Mountain, are given and range from about 6,550,000 long tons at Silver Lake to about 80,000 long tons of magnetite at Ship Mountain. The Vulcan deposit furnished 2,643,000 tons of ore to the Fontana steel mills during the period 1942–1944. Several of the other deposits have been worked in a small way for special-purpose ores.

The iron-ore deposits, except for those at Minarets, at Ship Mountain, and at Lake Hawley and Spencer Lakes, are typical contact metamorphic replacement bodies in limestones and dolomites. The ores in all the deposits are magnetite, except at Eagle Mountains, Kingston Range, and Ship Mountain, where the ores are mixtures of hematite and magnetite. At Minarets the magnetite ores replace metaandesite and are not related to any known intrusive mass. The small deposits around Lake Hawley and Spencer Lakes are described as replacement bodies in clastic sediments, dolomites, tuffs, and lamprophyric dikes. They are genetically related to an intrusive mass of metadiorite and its associated dikes. The ores at Ship Mountain form lenses in brecciated igneous and metamorphic rocks. Gypsum is present in the outcrops of several of the deposits, and in parts of several ore bodies the sulfur (pyrite) content is high for direct smelting. Phosphorus is low in all the analyses given.

On the surface the Kingston Range deposits appear to be among the largest and most promising of California; but, from diamond drilling done in 1924 and from later detailed geologic mapping, it is concluded that the ores lie in the overthrust block of a large thrust fault and that they do not continue into the underlying rocks.

The titaniferous magnetite bodies of the San Gabriel Mountains are in gabbro and anorthosite. The ores range from low grade to nearly solid titaniferous magnetite. In part of the rock the ilmenite and magnetite are minor accessory minerals, but they increase gradually in amount until some bodies are composed entirely of an intergrowth of magnetite and ilmenite. Thirty-six deposits were mapped and described. The best of the bodies contains about 250,000 tons of mineral, carrying 11–24 per cent of $TiO_2$. Details of a few placer deposits, found in the streams below the igneous masses, are given.

This book is mainly descriptive, but it brings together a great deal of miscellaneous information about the iron-ore deposits of California. It is a worth-while contribution and will be of special interest to anybody who desires to obtain impartial information about the character and the reserves of California iron-ore deposits.

CHARLES F. PARK, JR.

# COMMUNICATIONS AND ANNOUNCEMENTS

## GRANTS-IN-AID AVAILABLE FOR SCIENTIFIC RESEARCH

The American Academy of Arts and Sciences announces that income from its Permanent Science Fund will be disbursed as grants-in-aid in support of research projects in the fields of scientific business management, manufacture and commerce, engineering, psychology, education, economics and sociology, mathematics, physics, chemistry, astronomy, geology, geography, zoölogy, botany, anthropology, history and philology, medicine, surgery, agriculture, or any other science of any nature or description.

Applications for grants-in-aid are receivable on multiple forms, which will be supplied upon request to the chairman of the Permanent Science Fund Committee and are considered by the Committee on March 1 and October 1.

Requests for further information about conditions governing the grants should be addressed to: Hudson Hoagland, Chairman, Permanent Science Fund Committee, Worcester Foundation for Experimental Biology, 222 Maple Avenue, Shrewsbury, Massachusetts.

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## ERRATA

In "The Geomorphic History of the Carlsbad Caverns Area, New Mexico," by Leland Horberg, p. 472, legend for plate 1: reverse *B* and *C*.

In "The Ogallala Formation West of the Llano Estacado," by J Harlen Bretz and Leland Horberg, p. 492, legend for plate 1: reverse *B* and *C*.

# INDEX TO VOLUME 57


*Abundance of the rare gases in the earth and in the universe, The* (Hans E. Suess), 600

Acta Geologica Taiwanica, vol. 1, no. 1. Review by Ching Chang Woo, 437

*Africa, South, Some remarkable wind-faceted pebbles from near the mouth of the Umhlangakulu River, southern Natal* (Lester King), 413

*Age correlation, Leaching of carbonates in glacial drift and loess as a basis for* (Richard Foster Flint), 297

Ahlmann, Hans W:son. Glaciological research on the north Atlantic coast. Review by R. P. Sharp, 432

*(Ain), France, Rock-fragment slopes caused by past frost action in the Jura Mountains* (Sheldon Judson), 137

*Alaska, The frost-moved rubbles of Jumbo Dome and their significance in the Pleistocene chronology of* (Clyde Wahrhaftig), 216

*northern, Oriented lakes of* (Robert F. Black and William L. Barksdale), 105

*Thaw lakes and thaw sinks in the Imuruk Lake area, Seward Peninsula* (David M. Hopkins), 119

*Albite-water system in the gravitational field: a comment* (Hans Ramberg), 429

Amphibolite facies area (Hans Ramberg), 48

Anatexis, zone of (R. Perrin and M. Roubault), 362

*Ancient Arctica: remarks and supplement* (Wolf Maync), 314

*reply to remarks* (A. J. Eardley), 319

Ancient volcanoes of Oregon, The. By Howel Williams. Review by F. M. Byers, Jr., 324

*Antarctic ice specimens, The preservation of: a discussion* (Henri Bader), 427

*a reply* (Arthur David Howard), 428

Antevs, Ernst. Review of: Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit. By Hans Poser, 621

Review of: Die morphologischen Wirkungen des Eiszeitklimas im gletscherfreien Gebiet. By Julius Büdel, 235

Review of: Strukturböden, Solifluktion und Frostkilmate der Erde. By Carl Troll, 232

*Arctic Basin, Topography and sediments of the* (K. O. Emery), 512

*Arctica, Ancient: remarks and supplement* (Wolf Maync), 314

*reply to remarks* (A. J. Eardley), 319

Armstrong, H. S. Review of: Geological map of Manitoba, Map 850A, 323

Review of: Geological map of Saskatchewan, Map 895A, 321

Review of: Mineral map of Saskatchewan, Map 896A, 323

Arno Valley, stratigraphy of (Hallam L. Movius, Jr.), 385

*Ash, Great Plains Pleistocene volcanic, Source area of* (Ada Swineford), 307

*Australia, A review of tectonic relief in* (C. A. Cotton), 280

Bader, Henri. *The preservation of Antarctic ice specimens: a discussion*, 427

Bailey, Sturges W. *Liquid inclusions in granite thermometry*, 304

Balk, Robert. Review of: Emanuel Kayser's Abriss der Geologie. Vol. 2: Historische Geologie. By Roland Brinkmann, 441

Review of: The "Stratigraph," 327

Review of: Technische Gesteinskunde. By A. von Moos and F. de Quervain, 327

Barksdale, William L., and Robert F. Black. *Oriented lakes of northern Alaska*, 105

Barth, Tom. F. W. *The distribution of oxygen in the lithosphere: a reply*, 423

*Frequency distribution of the minerals in two petrographic provinces*, 55

*The use of oxygen as a reference substance in petrographic calculations: a supplement and a reply*, 425

Review of: Festschrift Paul Niggli, zu seinen 60. Geburtstag, 326

Review of: Geological explorations in western Tibet. By E. Norin, 326

Review of: Gesteine und Minerallagerstätten. Vol. 1: Allgemeine Lehre von den Gesteinen und Minerallagerstätten. By P. Niggli, 325

*Big Horn Mountains, Wyoming, Pleistocene ventifacts east of the* (Robert P. Sharp), 175

Black Cave (J Harlen Bretz), 457

*Black Hills, South Dakota, Petrogenesis of perthite pegmatites in the* (Riad A. Higazy), 555

Black, Robert F., and William L. Barksdale. *Oriented lakes of northern Alaska*, 105

Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit. By Hans Poser. Review by Ernst Antevs, 621

Bosteck, H. S. Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fifth parallel. Review by Leland Horberg, 445

*Boston Harbor, Recent shoreline changes at Shirley Gut* (Robert L. Nichols), 85

Botanical nomenclature, International rules of. Compiled by H. W. Camp, H. W. Rickett, and C. A. Weatherby. Review by E. C. Olson, 442

Bretz, J Harlen. *Carlsbad Caverns and other caves of the Guadalupe block, New Mexico*, 447

and Leland Horberg. *Caliche in southeastern New Mexico*, 491
and Leland Horberg. *The Ogallala formation west of the Llano Extacado*, 477
Brinkmann, Roland. Emanuel Kayser's Abriss der Geologie. Vol. 2: Historische Geologie. Review by Robert Balk, 441
*Briquetted coal, Naturally, from a bed in western Pennsylvania* (James M. Schopf), 79
Bryan, Kirk. *The geologic implications of cryopedology*, 101
Bryan, W. H. The geological approach to the study of soils. Review by C. A. Cotton, 98
Büdel, Julius. Die morphologischen Wirkungen des Eiszeitklimas im gletscherfreien Gebiet. Review by Ernst Antevs, 235
Bullen, K. E. An introduction to the theory of seismology. Review by M. King Hubbert, 435
*Bureau County, Illinois, A possible fossil ice wedge in* (Leland Horberg), 132
Byers, F. M., Jr. Review of: The ancient volcanoes of Oregon. By Howel Williams, 324

Cady, G. H. Review of: Die Metamorphose der Kohle und ihr Einfluss auf die sichtbaren derselben. By W. Petrascheck, 551
Cailleux, André, Études de cryopédologie. Review by Leland Horberg, 236
*Caliche, in southeastern New Mexico* (J Harlen Bretz and Leland Horberg), 491
age and climatic implications, 509
alteration of, 504
lithologic description of, 502
origin of, 507
profile, 493
*California Tertiary reservoir sands, Cementation of some* (Charles M. Gilbert), 1
Camp, H. W., H. W. Rickett, and C. A. Weatherby (comps.). International rules of botanical nomenclature. Review by E. C. Olson, 442
*Carbonates, Leaching of, in glacial drift and loess as a basis for age correlation* (Richard Foster Flint), 297
*Carlsbad Caverns area, New Mexico, Geomorphic history of the* (Leland Horberg), 464
*Carlsbad Caverns and other caves of the Guadalupe block, New Mexico* (J Harlen Bretz), 447
Carolina bays, comparison of the oriented lakes of northern Alaska with (Robert F. Black and William L. Barksdale), 115
Cavern development, time of, in Guadalupe Range (Leland Horberg), 475
*Caves, Carlsbad Caverns and other, of the Guadalupe block, New Mexico* (J Harlen Bretz), 447
*Cellular structures in glass as related to structures in lava* (E. F. Osborn), 73
*Cementation of some California Tertiary reservoir sands* (Charles M. Gilbert), 1
*Cementing agents, Sandstone types: their abundance and* (Sefton L. Tallman), 582
Chambers's mineralogical dictionary. Review by D. J. Fisher, 99
Chayes, F. *On ratio correlation in petrography*, 239
*Chemical stability, Radial diffusion and, in the gravitational field: a discussion* (Jean Verhoogen), 93
*a reply* (Hans Ramberg), 95
*Chronology, Pleistocene, of Alaska, The frost-moved rubbles of Jumbo Dome and their significance in the* (Clyde Wahrhaftig), 216
*Classification, The facies, of rocks: a clue to the origin of quartzo-feldspathic massifs and veins* (Hans Ramberg), 18
Climate, of the Arctic Coastal Plain (Robert F. Black and William L. Barksdale), 107
Climatic accidents in landscape making. By C. A. Cotton. Review by Leland Horberg, 238
Climatic implications of caliche (J Harlen Bretz and Leland Horberg), 509
*Coal, Naturally briquetted, from a bed in western Pennsylvania* (James M. Schopf), 79
Conybeare, C. E. B. *Stylolites in pre-Cambrian quartzite*, 83
*Correlation, age, Leaching of carbonates in glacial drift and loess as a basis for* (Richard Foster Flint), 297
*On ratio, in petrography* (F. Chayes), 239
Cotton, C. A. Climatic accidents in landscape making. Review by Leland Horberg, 238
*A review of tectonic relief in Australia*, 280
Review of: The geological approach to the study of soils. By W. H. Bryan, 98
Review of: Landscape study in South Africa. By Lester C. King, 236
Cottonwood Cave (J Harlen Bretz), 456
Cretaceous and Tertiary subsurface geology: the stratigraphy, paleontology, and sedimentology of three deep test wells on the eastern shore of Maryland. Review by W. C. Krumbein, 620
*Crustal movement in the Great Lakes area: a discussion* (William Herbert Hobbs), 312
*Cryopedology, The geologic implications of* (Kirk Bryan), 101
Crystallization, evolution of (R. Perrin and M. Roubault), 370

Differentiation (R. Perrin and M. Roubault), 360
*Diffusion, Radial, and chemical stability in the gravitational field: a discussion* (Jean Verhoogen), 93
*a reply* (Hans Ramberg), 95
Diffusion and reactions in the solid state, evidence of origin of granite by (R. Perrin and M. Roubault), 364
*Distribution of oxygen in the lithosphere, The, and Oxygen in rocks: a basis for petrographic calculations: a discussion* (Ivan Th. Rosenqvist), 420
*a reply* (Tom. F. W. Barth), 423
*comments on Barth's reply* (Ivan Th. Rosenqvist), 425
Drainage, of the Arctic Coastal Plain (Robert F. Black and William L. Barksdale), 106

evolution, in Guadalupe Range (Leland Horberg), 474
*Drift, glacial, and loess, Leaching of carbonates in, as a basis for age correlation* (Richard Foster Flint), 297
*Driftless Area of southern Wisconsin, Periglacial features in the* (H. T. U. Smith), 196

Eardley, A. J. *Ancient Arctica: reply to remarks*, 319
Review of: Symposium of information relative to uses of aerial photographs by geologists. Compiled by H. T. U. Smith, 439
*Earth and in the universe, The abundance of the rare gases in the* (Hans E. Suess), 600
Earthquake epicenters, in Arctic region (K. O. Emery), 515
*Edelgase, Die Häufigkeit der, auf der Erde und im Kosmos* (Hans E. Suess), 600
Emery, K. O. *Topography and sediments of the Arctic Basin*, 512
Emplacement of granite (R. Perrin and M. Roubault), 361
Enstatite-orthoferrosilite series (Th. G. Sahama and D. R. Torgeson), 258
Epidote-amphibolite facies area (Hans Ramberg), 49
*Equating of volume and weight, Replacement and the* (John Ridge), 522
*Erde und im Kosmos, Die Häufigkeit der Edelgase auf der* (Hans E. Suess), 600
Études de cryopédologie. By André Cailleux. Review by Leland Horberg, 236
*Europe, southern and southwestern, Villafranchian stratigraphy in* (Hallam L. Movius, Jr.), 380

*Facies classification of rocks, The: a clue to the origin of quartzo-feldspathic massifs and veins* (Hans Ramberg), 18
*Feldspars, Recent results of investigations on the* (Alexander Köhler), 592
Festschrift Paul Niggli, zu seinen 60. Geburtstag. Review by Tom. F. W. Barth, 326
Fisher, D. J. Review of: Chambers's mineralogical dictionary, 99
Flint, Richard Foster. *Leaching of carbonates in glacial drift and loess as a basis for age correlation*, 297
Chester R. Longwell, and Adolph Knopf. Physical geology. Review by Lou Williams Page, 623
*Florida landslide, A* (Richard H. Jordan), 418
Forsterite-fayalite series (Th. G. Sahama and D. R. Torgeson), 256
*Fossil ice wedge, A possible, in Bureau County, Illinois* (Leland Horberg), 132
*France, Rock-fragment slopes caused by past frost action in the Jura Mountains (Ain)* (Sheldon Judson), 137
stratigraphy of Saône Basin (Hallam L. Movius, Jr.), 393
*Frequency distribution of the minerals in two petrographic provinces* (Tom. F. W. Barth), 55

Friedman, I. I. *A proposed method for the measurement of geologic temperatures*, 618
Frost action, intensive, the study of (Kirk Bryan), 101
*past, Rock-fragment slopes caused by, in the Jura Mountains (Ain), France* (Sheldon Judson), 137
*Periglacial features in the Driftless Area of southern Wisconsin* (H. T. U. Smith), 196
*Some periglacial features in central Montana* (J. P. Schafer), 154
*Stone nets, stone stripes, and soil stripes in the Wind River Mountains, Wyoming* (Gerald M. Richmond), 143
*Frost-moved rubbles of Jumbo Dome, The, and their significance in the Pleistocene chronology of Alaska* (Clyde Wahrhaftig), 216
Frost riving and soil formation (David M. Hopkins), 120
Fundamentals of physical science. By Konrad Bates Krauskopf. Review by Thornton Page, 438

Geikielite and ilmenite, stability relation of (Th. G. Sahama and D. R. Torgeson), 260
*Geochemical researches in Japan* (Kazuo Yamasaki), 414
Geochemistry. *Albite-water system in the gravitational field: a comment* (Hans Ramberg), 429
*The distribution of oxygen in the lithosphere, and, Oxygen in rocks: a basis for petrographic calculations: a discussion* (Ivan Th. Rosenqvist), 420
*a reply* (Tom. F. W. Barth), 423
*comments on Barth's reply* (Ivan Th. Rosenqvist), 425
*The facies classification of rocks: a clue to the origin of quartzo-feldspathic massifs and veins* (Hans Ramberg), 18
*Liquid inclusions in granite thermometry* (Sturges W. Bailey), 304
*Oxidation and reduction in* (Brian Mason), 62
*Radial diffusion and chemical stability in the gravitational field: a discussion* (Jean Verhoogen), 93
*a reply* (Hans Ramberg), 95
*Replacement and the equating of volume and weight* (John Ridge), 522
*Some examples of the application of thermochemistry to petrology* (Th. G. Sahama and D. R. Torgeson), 255
*of thallium, A note on the* (Kalervo Rankama), 608
*The use of oxygen as a reference substance in petrographic calculations: a supplement and a reply* (Tom. F. W. Barth), 425
*Geologic implications of cryopedology, The* (Kirk Bryan), 101
*Geologic temperatures, A proposed method for the measurement of* (I. I. Friedman), 618
Geological approach to the study of soils, The. By W. H. Bryan. Review by C. A. Cotton, 98

Geological explorations in western Tibet. By E. Norin. Review by Tom. F. W. Barth, 326
Geological map of Manitoba, Map 850A. Review by H. S. Armstrong, 323
Geological map of the Province of Ontario: Map no. 1946-3. Compiled by J. Satterly. Review by F. J. Pettijohn, 321
Geological map of Saskatchewan, Map 895A. Review by H. S. Armstrong, 321
Geology of the Arctic Coastal Plain (Robert F. Black and William L. Barksdale), 108
*Geomorphic history of the Carlsbad Caverns area, New Mexico* (Leland Horberg), 464
Geomorphology. *A review of tectonic relief in Australia* (C. A. Cotton), 280
Geophysics. *Isostasy—extended* (Ross Gunn), 263
Gesteine und Minerallagerstätten. Vol. 1: Allgemeine Lehre von den Gesteinen und Minerallagerstätten. By P. Niggli. Review by Tom. F. W. Barth, 325
Gilbert, Charles M. *Cementation of some California Tertiary reservoir sands*, 1
*Glacial drift and loess, Leaching of carbonates in, as a basis for age correlation* (Richard Foster Flint), 297
Glacial erosion in Arctic region (K. O. Emery), 514
Glaciation in Jumbo Dome area (Clyde Wahrhaftig), 219
Glaciological research on the north Atlantic coast. By Hans W:son Ahlmann. Review by R. P. Sharp, 432
*Glass, Cellular structures in, as related to structures in lava* (E. F. Osborn), 73
*Granite problem, On the* (R. Perrin and M. Roubault), 357
*Granite thermometry, Liquid inclusions in* (Sturges W. Bailey), 304
Granitization, rate of (R. Perrin and M. Roubault), 370
Granulite facies area (Hans Ramberg), 35
*Gravitational field, Albite-water system in the: a comment* (Hans Ramberg), 429
*Radial diffusion and chemical stability in the: a discussion* (Jean Verhoogen), 93
*a reply* (Hans Ramberg), 95
*Great Lakes area, Crustal movement in the: a discussion* (William Herbert Hobbs), 312
*Great Plains Pleistocene volcanic ash, Source area of* (Ada Swineford), 307
Greenland. *Ancient Arctica: remarks and supplement* (Wolf Mayne), 314
*Guadalupe block, New Mexico, Carlsbad Caverns and other caves of the* (J Harlen Bretz), 447
Guadalupe Range. *Geomorphic history of the Carlsbad Caverns area, New Mexico* (Leland Horberg), 464
Gunn, Ross. *Isostasy—extended*, 263

*Häufigkeit der Edelgase auf der Erde und im Kosmos, Die* (Hans E. Suess), 600
Hagner, A. F. Review of: Mineral resources of Colorado. Prepared under supervision of John W. Vanderwilt, 439
Heat conduction, with engineering and geological applications. By L. R. Ingersoll, O. J. Zobel, and A. C. Ingersoll. Review by T. S. Lovering, 327
Hidden Cave (J Harlen Bretz), 457
Higazy, Riad A. *Petrogenesis of perthite pegmatites in the Black Hills, South Dakota*, 555
Historical geology. By Russell C. Hussey. Review by Robert L. Miller, 437
*History, Geomorphic, of the Carlsbad Caverns area, New Mexico* (Leland Horberg), 464
Hobbs, William Herbert. *Crustal movement in the Great Lakes area: a discussion*, 312
Hohlt, Richard B. The nature and origin of limestone porosity. Review by F. A. Johnson, Jr., 441
Hopkins, David M. *Thaw lakes and thaw sinks in the Imuruk Lake area, Seward Peninsula, Alaska*, 119
Horberg, Leland *Geomorphic history of the Carlsbad Caverns area, New Mexico*, 464
*A possible fossil ice wedge in Bureau County, Illinois*, 132
Review of: Climatic accidents in landscape making. By C. A. Cotton, 238
Review of: Études de cryopédologie. By André Cailleux, 236
Review of: Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fifth parallel. By H. S. Bosteck, 445
and J Harlen Bretz. *Caliche in southeastern New Mexico*, 491
and J Harlen Bretz. *The Ogallala formation west of the Llano Estacado*, 477
Howard, Arthur David. *The preservation of Antarctic ice specimens: a reply*, 428
Hubbert, M. King. Review of: An introduction to the theory of seismology. By K. E. Bullen, 435
Hussey, Russell C. Historical geology. Review by Robert L. Miller, 437

*Ice specimens, Antarctic, The preservation of: a discussion* (Henri Bader), 427
*a reply* (Arthur David Howard), 428
*Ice wedge, A possible fossil, in Bureau County, Illinois* (Leland Horberg), 132
*Illinois, A possible fossil ice wedge in Bureau County* (Leland Horberg), 132
*Trivoli sandstone of Williamson County* (Raymond Siever), 614
Ilmenite and geikielite, stability relation of (Th. G. Sahama and D. R. Torgeson), 260
*Imuruk Lake area, Seward Peninsula, Alaska, Thaw lakes and thaw sinks in the* (David M. Hopkins), 119
*Inclusions, Liquid, in granite thermometry* (Sturges W. Bailey), 304
*liquid, Structural petrology of planes of* (O. F. Tuttle), 331

Ingersoll, A. C., L. R. Ingersoll, and O. J. Zobel. Heat conduction, with engineering and geological applications. Review by T. S. Lovering, 327
Ingersoll, L. R., O. J. Zobel, and A. C. Ingersoll. Heat conduction, with engineering and geological applications. Review by T. S. Lovering, 327
Intra-Pacific rock province (Tom. F. W. Barth), 57
*Investigations on the feldspars, Recent results of* (Alexander Köhler), 592
Involutions, description and origin (J. P. Schafer), 156, 160
Iron resources of California, The. Review by Charles F. Park, Jr., 626
Isobary (Ross Gunn), 268
*Isostasy—extended* (Ross Gunn), 263
Italy, southern, stratigraphy in (Hallam L. Movius, Jr.), 389

*Japan, Geochemical researches in* (Kazuo Yamasaki), 414
Johnson, F. A., Jr. Review of: The nature and origin of limestone porosity. By Richard B. Hohlt, 441
Jordan, Richard H. *A Florida landslide*, 418
Judson, Sheldon. *Rock-fragment slopes caused by past frost action in the Jura Mountains (Ain), France*, 137
*Jumbo Dome, The frost-moved rubbles of, and their significance in the Pleistocene chronology of Alaska* (Clyde Wahrhaftig,) 216
*Jura Mountains (Ain), France, Rock-fragment slopes caused by past frost action in the* (Sheldon Judson), 137

Kayser's, Emanuel, Abriss der Geologie. Vol. 2: Historische Geologie. By Roland Brinkmann. Review by Robert Balk, 441
King, Lester C. Landscape study in South Africa. Review by C. A. Cotton, 236
*Some remarkable wind-faceted pebbles from near the mouth of the Umhlangakulu River, southern Natal, South Africa*, 413
Knopf, Adolph, Chester R. Longwell, and Richard F. Flint. Physical geology. Review by Lou Williams Page, 623
Köhler, Alexander. *Recent results of investigations on the feldspars*, 592
*Kosmos, Die Häufigkeit der Edelgase auf der Erde und im* (Hans E. Suess), 600
Krauskopf, Konrad Bates. Fundamentals of physical science. Review by Thornton Page, 438
Krumbein, W. C. Review of: Cretaceous and Tertiary subsurface geology: the stratigraphy, paleontology, and sedimentology of three deep test wells on the eastern shore of Maryland, 620

*Lakes, Oriented, of northern Alaska* (Robert F. Black and William L. Barksdale), 105
*Thaw, and thaw sinks in the Imuruk Lake area, Seward Peninsula, Alaska* (David M. Hopkins), 119
Landscape study in South Africa. By Lester C. King. Review by C. A. Cotton, 236
*Landslide, A Florida* (Richard H. Jordan), 418
*Lava, Cellular structures in glass as related to structures in* (E. F. Osborn), 73
*Leaching of carbonates in glacial drift and loess as a basis for age correlation* (Richard Foster Flint), 297
Limestone porosity, The nature and origin of. By Richard B. Hohlt. Review by F. A. Johnson, Jr., 441
*Liquid inclusions in granite thermometry* (Sturges W. Bailey), 304
*Liquid inclusions, Structural petrology of planes of* (O. F. Tuttle), 331
Lithologic description of caliche (J Harlen Bretz and Leland Horberg), 502
*Lithosphere, The distribution of oxygen in the, and, Oxygen in rocks: a basis for petrographic calculations: a discussion* (Ivan Th. Rosenqvist), 420
*a reply* (Tom. F. W. Barth), 423
*comments on Barth's reply* (Ivan Th. Rosenqvist), 425
*Llano Estacado, The Ogallala formation west of the* (J Harlen Bretz and Leland Horberg), 477
*Loess, Leaching of carbonates in glacial drift and, as a basis for age correlation* (Richard Foster Flint), 297
Longwell, Chester R., Adolph Knopf, and Richard F. Flint, Physical geology. Review by Lou Williams Page, 623
Lovering, T. S. Review of: Heat conduction, with engineering and geological applications. By L. R. Ingersoll, O. J. Zobel, and A. C. Ingersoll, 327

McKinstry, H. E. Mining geology. Review by Olaf N. Rove, 625
McKittrick Cave (J Harlen Bretz), 459
Magmatic theory versus the metamorphic-metasomatic theory (Hans Ramberg), 27
"Magmatist" hypotheses in origin of granite (R. Perrin and M. Roubault), 358
Mason, Brian. *Oxidation and reduction in geochemistry*, 62
Massachusetts. *Recent shoreline changes at Shirley Gut, Boston Harbor* (Robert L. Nichols), 85
*Massifs and veins, quartzo-feldspathic, The facies classification of rocks: a clue to the origin of* (Hans Ramberg), 18
Mather, Kirtley F. Review of: The religion of the modern scientist (neo-materialism). By S. W. Tromp, 625
Mayne, Wolf. *Ancient Arctica: remarks and supplement*, 314
*Measurement of geologic temperatures, A proposed method for the* (I. I. Friedman), 618
Metamorphic-metasomatic theory, magmatic theory versus the (Hans Ramberg), 27

Metamorphism in folded mountains (Hans Ramberg), 24
Metamorphose der Kohle, Die, und ihr Einfluss auf die sichtbaren derselben. By W. Petrascheck. Review by G. H. Cady, 551
*Mid-Atlantic Ridge, Rocks of the* (S. J. Shand), 89
Miller, Robert L. Review of: Historical geology. By Russell C. Hussey, 437
Mineral map of Saskatchewan, Map 896A. Review by H. S. Armstrong, 323
Mineral resources of Colorado. Prepared under supervision of John W. Vanderwilt. Review by A. F. Hagner, 439
Mineralogy of California Tertiary sandstones (Charles M. Gilbert), 2
*Minerals, Frequency distribution of, in two petrographic provinces* (Tom. F. W. Barth), 55
Mining geology. By H. E. McKinstry. Review by Olaf N. Rove, 625
*Montana, central, Some periglacial features in* (J. P. Schafer), 154
Moos, A. von, and F. de Quervain. Technische Gesteinskunde. Review by Robert Balk, 327
Morphologischen Wirkungen des Eiszeitklimas im gletscherfreien Gebiet, Die. By Julius Büdel. Review by Ernst Antevs, 235
*Movement, Crustal, in the Great Lakes area: a discussion* (William Herbert Hobbs), 312
Movius, Hallam L., Jr. *Villafranchian stratigraphy in southern and southwestern Europe*, 380
Mudgett's Cave (J Harlen Bretz), 459

*Natal, southern, South Africa, Some remarkable wind-faceted pebbles from near the mouth of the Umhlangakulu River* (Lester King), 413
*Nets, Stone, stone stripes, and soil stripes in the Wind River Mountains, Wyoming* (Gerald M. Richmond), 143
New Cave (J Harlen Bretz), 454
*New Mexico, Carlsbad Caverns and other caves of the Guadalupe block* (J Harlen Bretz), 447
*Geomorphic history of the Carlsbad Caverns area* (Leland Horberg), 464
*The Ogallala formation west of the Llano Estacado* (J Harlen Bretz and Leland Horberg), 477
*Source area of the Great Plains Pleistocene volcanic ash* (Ada Swineford), 307
*southeastern, Caliche in* (J Harlen Bretz and Leland Horberg), 491
New South Wales (C. A. Cotton), 280
Nichols, Robert L. *Recent shoreline changes at Shirley Gut, Boston Harbor*, 85
Niggli, P. Gesteine und Minerallagerstätten. Vol. 1: Allgemeine Lehre von den Gesteinen und Minerallagerstätten. Review by Tom. F. W. Barth, 325
Norin, E. Geological explorations in western Tibet. Review by Tom. F. W. Barth, 326

Oceanographical Institute of Taiwan, Bulletin of the, nos. 1–3. Review by Ching Chang Woo, 436
*Ogallala formation west of the Llano Estacado, The* (J Harlen Bretz and Leland Horberg), 477
Olson, E. C. Review of: International rules of botanical nomenclature. Compiled by H. W. Camp, H. W. Rickett, and C. A. Weatherby, 442
Review of: The Permo-Triassic formations. By R. L. Sherlock, 443
*Oriented lakes of northern Alaska* (Robert F. Black and William L. Barksdale), 105
Orthoferrosilite, stability of (Th. G. Sahama and D. R. Torgeson), 259
Osborn, E. F. *Cellular structures in glass as related to structures in lava*, 73
Oslo petrographic province (Tom. F. W. Barth), 57
*Oxidation and reduction in geochemistry* (Brian Mason), 62
*Oxygen in the lithosphere, The distribution of, and Oxygen in rocks: a basis for petrographic calculations: a discussion* (Ivan Th. Rosenqvist), 420
*a reply* (Tom. F. W. Barth), 423
*comments on Barth's reply* (Ivan Th. Rosenqvist), 425
*Oxygen, The use of, as a reference substance in petrographic calculations: a supplement and a reply* (Tom. F. W. Barth), 425

Page, Lou Williams. Review of: Physical geology. By Chester R. Longwell, Adolph Knopf, and Richard F. Flint, 623
Page, Thornton. Review of: Fundamentals of physical science. By Konrad Bates Krauskopf, 438
Park, Charles F., Jr. Review of: The iron resources of California, 626
Pearlette ash. *Source area of Great Plains Pleistocene volcanic ash* (Ada Swineford), 307
Pearson's general formula for ratio correlation (F. Chayes), 240
*Pebbles, Some remarkable wind-faceted, from near the mouth of the Umhlangakulu River, southern Natal, South Africa* (Lester King), 413
Peck, Ralph B., and Karl Terzaghi. Soil mechanics in engineering practice. Review by Parker D. Trask, 622
Pecos lowland, valley fill in the (Leland Horberg), 468
Pegmatites, conclusions regarding the formation of (Hans Ramberg), 50
*perthite, in the Black Hills, South Dakota, Petrogenesis of* (Riad A. Higazy), 555
*Pennsylvania, western, Naturally briquetted coal from a bed in* (James M. Schopf), 79
Periglacial climate (Kirk Bryan), 2
(J. P. Schafer), 171
*Periglacial features in central Montana, Some* (J. P. Schafer), 154
*Periglacial features in the Driftless Area of southern Wisconsin* (H. T. U. Smith), 196
Permafrost. *Oriented lakes of northern Alaska* (Robert F. Black and William L. Barksdale), 105
*A possible fossil ice wedge in Bureau County, Ill.*

*nois* (Leland Horberg), 132
*Thaw lakes and thaw sinks in the Imuruk Lake area, Seward Peninsula, Alaska* (David M. Hopkins), 119
Permo-Triassic formations, The. By R. L. Sherlock. Review by E. C. Olson, 443
Perrin, R., and M. Roubault. *On the granite problem*, 357
*Perthite pegmatites in the Black Hills, South Dakota, Petrogenesis of* (Riad A. Higazy), 555
Petrascheck, W. Die Metamorphose der Kohle und ihr Einfluss auf die sichtbaren derselben. Review by G. H. Cady, 451
*Petrogenesis of perthite pegmatites in the Black Hills, South Dakota* (Riad A. Higazy), 555
*Petrographic calculations, Oxygen in rocks: a basis for, and, The distribution of oxygen in the lithosphere: a discussion* (Ivan Th. Rosenqvist), 420
*The use of oxygen as a reference substance in: a supplement and a reply* (Tom. F. W. Barth), 425
*Petrographic provinces, two, Frequency distribution of the minerals in* (Tom. F. W. Barth), 55
Petrography, of Pearlette tuff and pumice (Ada Swineford), 309
*On ratio correlation in* (F. Chayes), 239
*Rocks of the Mid-Atlantic Ridge* (S. J. Shand), 89
Petroleum Technology, Journal of, vol. 1, no. 1. Review by F. J. Pettijohn, 442
Petrology. *Cementation of some California Tertiary reservoir sands* (Charles M. Gilbert), 1
*The facies classification of rocks: a clue to the origin of quartzo-feldspathic massifs and veins* (Hans Ramberg), 18
*Recent results of investigations on the feldspars* (Alexander Köhler), 592
*Some examples of the application of thermochemistry to* (Th. G. Sahama and D. R. Torgeson), 255
*Structural, of planes of liquid inclusions* (O. F. Tuttle), 331
Pettijohn, F. J. Review of: Geological map of the Province of Ontario: Map no. 1946-3. Compiled by J. Satterly, 321
Review of: Journal of Petroleum Technology, vol. 1, no. 1, 442
Physical geology. By Chester R. Longwell, Adolph Knopf, and Richard F. Flint. Review by Lou Williams Page, 623
Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fifth parallel. By H. S. Bosteck. Review by Leland Horberg, 445
Piedmont plains and terraces in Guadalupe Range (Leland Horberg), 470
*Planes of liquid inclusions, Structural petrology of* (O. F. Tuttle), 331
*Pleistocene chronology of Alaska, The frost-moved rubbles of Jumbo Dome and their significance in the* (Clyde Wahrhaftig), 216
*Pleistocene ventifacts east of the Big Horn Mountains, Wyoming* (Robert P. Sharp), 175
*Pleistocene volcanic ash, Great Plains, Source area of* (Ada Swineford), 307
Pliocene. *The Ogallala formation west of the Llano Estacado* (J Harlen Bretz and Leland Horberg), 477
Po Valley, stratigraphy of (Hallam L. Movius, Jr.), 381
Poser, Hans. Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit. Review by Ernst Antevs, 621
*Pre-Cambrian quartzite, Stylolites in* (C. E. B. Conybeare), 83
*Preservation of Antarctic ice specimens, The: a discussion* (Henri Bader), 427
*a reply* (Arthur David Howard), 428
Puy-en-Velay Basin, stratigraphy of (Hallam L. Movius, Jr.), 400

*Quartzite, pre-Cambrian, Stylolites in* (C. E. B. Conybeare), 83
*Quartzo-feldspathic massifs and veins, The facies classification of rocks: a clue to the origin of* (Hans Ramberg), 18
Quartzo-feldspathic rocks, position in the mineral facies scheme (Hans Ramberg), 47
Queensland (C. A. Cotton), 287
Quervain, F. de, and A. von Moos. Technische Gesteinskunde. Review by Robert Balk, 327

*Radial diffusion and chemical stability in the gravitational field: a discussion* (Jean Verhoogen), 93
*a reply* (Hans Ramberg), 95
Rainfall effects. *A Florida landslide* (Richard H. Jordan), 418
Ramberg, Hans. *Albite-water system in the gravitational field: a comment*, 429
*The facies classification of rocks: a clue to the origin of quartzo-feldspathic massifs and veins*, 18
*Radial diffusion and chemical stability in the gravitational field: a reply*, 95
Rankama, Kalervo. *A note on the geochemistry of thallium*, 608
*Rare gases, The abundance of, in the earth and in the universe* (Hans E. Suess), 600
*Ratio correlation in petrography, On* (F. Chayes), 239
*Reduction, Oxidation and, in geochemistry* (Brian Mason), 62
*Relief, tectonic, in Australia, a review of* (C. A. Cotton), 280
Religion of the modern scientist (neo-materialism), The. By S. W. Tromp. Review by Kirtley F. Mather, 625
*Replacement and the equating of volume and weight* (John Ridge), 522
*Researches, Geochemical, in Japan* (Kazuo Yamasaki), 414

Richmond, Gerald M. *Stone nets, stone stripes, and soil stripes in the Wind River Mountains, Wyoming*, 143
Rickett, H. W., H. W. Camp, and C. A. Weatherby, (comps.). International rules of botanical nomenclature. Review by E. C. Olson, 442
Ridge, John. *Replacement and the equating of volume and weight*, 522
Riley, N. A. Review of: Sequence in layered rocks. By Robert R. Shrock, 99
*Rock-fragment slopes caused by past frost action in the Jura Mountains (Ain), France* (Sheldon Judson), 137
*Rocks, The facies classification of: a clue to the origin of quartzo-feldspathic massifs and veins* (Hans Ramberg), 18
*of the Mid-Atlantic Ridge* (S. J. Shand), 89
*Oxygen in: a basis for petrographic calculations, and, The distribution of oxygen in the lithosphere: a discussion* (Ivan Th. Rosenqvist), 420
Rosenqvist, Ivan Th. *The distribution of oxygen in the lithosphere, and, Oxygen in rocks: a basis for petrographic calcultaions: a discussion*, 420
*comments on Barth's reply*, 425
Roubault, M., and R. Perrin. *On the granite problem*, 357
Rove, Olaf N. Review of: Mining geology. By H. E. McKinstry, 625
*Rubbles, The frost-moved, of Jumbo Dome and their significance in the Pleistocene chronology of Alaska* (Clyde Wahrhaftig), 216

Sahama, Th. G., and D. R. Torgeson. *Some examples of the application of thermochemistry to petrology*, 255
*Sands, California Tertiary reservoir, Cementation of some* (Charles M. Gilbert), 1
*Sandstone, Trivoli, of Williamson County, Illinois* (Raymond Siever), 614
*types: their abundance and cementing agents* (Sefton L. Tallman), 582
Sandstones, classification of (Sefton L. Tallman), 583
Saône Basin, France, stratigraphy of (Hallam L. Movius, Jr.), 393
Saskatchewan. *Stylolites in pre-Cambrian quartzite* (C. E. B. Conybeare), 83
Satterly, J., comp. Geological map of the Province of Ontario: Map no. 1946-3. Review by F. J. Pettijohn, 321
Schafer, J. P. *Some periglacial features in central Montana*, 154
Schopf, James M. *Naturally briquetted coal from a bed in western Pennsylvania*, 79
*Sediments, Topography and, of the Arctic Basin* (K. O. Emery), 512
Seismology, An introduction to the theory of. By K. E. Bullen. Review by M. King Hubbert, 435
Sequence in layered rocks. By Robert R. Shrock. Review by N. A. Riley, 99
*Seward Peninsula, Alaska, Thaw lakes and thaw sinks in the Imuruk Lake area* (David M. Hopkins), 119
Shand, S. J. *Rocks of the Mid-Atlantic Ridge*, 89
Sharp, Robert P. *Pleistocene ventifacts east of the Big Horn Mountains, Wyoming*, 175
Review of: Glaciological research on the north Atlantic coast. By Hans W:son Ahlmann, 432
Shepard, Francis P. Submarine geology. Review by H. C. Stetson, 443
Sherlock, R. L. The Permo-Triassic formations. Review by E. C. Olson, 443
*Shirley Gut, Boston Harbor, Recent shoreline changes at* (Robert L. Nichols), 85
*Shoreline changes, Recent, at Shirley Gut, Boston Harbor* (Robert L. Nichols), 85
Shrock, Robert R. Sequence in layered rocks. Review by N. A. Riley, 99
Sicily, stratigraphy of (Hallam L. Movius, Jr.), 389
Siever, Raymond. *Trivoli sandstone of Williamson County, Illinois*, 614
*Sinks, thaw, Thaw lakes and, in the Imuruk Lake area, Seward Peninsula, Alaska* (David M. Hopkins), 119
*Slopes, Rock-fragment, caused by past frost action in the Jura Mountains (Ain), France* (Sheldon Judson), 137
Smith, H. T. U. *Periglacial features in the Driftless Area of southern Wisconsin*, 196
comp. Symposium of information relative to uses of aerial photographs by geologists. Review by A. J. Eardley, 439
Soil formation, frost riving and (David M. Hopkins), 120
Soil mechanics in engineering practice. By Karl Terzaghi and Ralph B. Peck. Review by Parker D. Trask, 622
*Soil stripes, Stone nets, stone stripes, and, in the Wind River Mountains, Wyoming* (Gerald M. Richmond), 143
Solution work. *Carlsbad Caverns and other caves of* the *Guadalupe block, New Mexico* (J Harlen Bretz), 447
Soundings in Arctic Basin (K. O. Emery), 512
*South Africa, Some remarkable wind-faceted pebbles from near the mouth of the Umhlangakulu River, southern Natal* (Lester King), 413
South Australia (C. A. Cotton), 293
*South Dakota, Petrogenesis of perthite pegmatites in the Black Hills* (Riad A. Higazy), 555
*Stability, chemical, Radial diffusion and, in the gravitational field: a discussion* (Jean Verhoogen), 93
*a reply* (Hans Ramberg), 95
Stetson, H. C. Review of: Submarine geology. By Francis P. Shepard, 443
*Stone nets, stone stripes, and soil stripes in the Wind River Mountains, Wyoming* (Gerald M. Richmond), 143
*Stone stripes, Stone nets, and soil stripes in the Wind*

*River Mountains, Wyoming* (Gerald M. Richmond), 143
"Stratigraph," The. Review by Robert Balk, 327
*Stratigraphy, Villafranchian, in southern and southwestern Europe* (Hallam L. Movius, Jr.), 380
*Stripes, soil, Stone nets, stone stripes, and, in the Wind River Mountains, Wyoming* (Gerald M. Richmond), 143
*Stripes, stone, Stone nets, and soil stripes in the Wind River Mountains, Wyoming* (Gerald M. Richmond), 143
*Structural petrology of planes of liquid inclusions* (O. F. Tuttle), 331
*Structures, Cellular, in glass as related to structures in lava* (E. F. Osborn), 73
Strukturböden, Solifluktion und Frostklimate der Erde. By Carl Troll. Review by Ernst Antevs, 232
*Stylolites in pre-Cambrian quartzite* (C. E. B. Conybeare), 83
Submarine geology. By Francis P. Shepard. Review by H. C. Stetson, 443
Suess, Hans E. *Die Häufigkeit der Edelgase auf der Erde und im Kosmos*, 600
Swineford, Ada. *Source area of Great Plains Pleistocene volcanic ash*, 307
Symposium of information relative to uses of aerial photographs by geologists. Compiled by H. T. U. Smith. Review by A. J. Eardley 439

Tallman, Sefton L. *Sandstone types: their abundance and cementing agents*, 582
Tasmania (C. A. Cotton), 291
Techniques and methods. *The preservation of Antarctic ice specimens: a discussion* (Henri Bader), 427
*a reply* (Arthur David Howard), 428
Technische Gesteinskunde. By A. von Moos and F. de Quervain. Review by Robert Balk, 327
*Tectonic relief in Australia, A review of* (C. A. Cotton), 280
*Temperatures, geologic, A proposed method for the measurement of* (I. I. Friedman), 618
Terraces, piedmont plains and, in the Guadalupe Range (Leland Horberg), 470
*Tertiary reservoir sands, California, Cementation of some* (Charles M. Gilbert), 1
Terzaghi, Karl, and Ralph B. Peck. Soil mechanics in engineering practice. Review by Parker D. Trask, 622
Texture of California Tertiary sandstones (Charles M. Gilbert), 12
*Thallium, A note on the geochemistry of* (Kalervo Rankama), 608
*Thaw lakes and thaw sinks in the Imuruk Lake area, Seward Peninsula, Alaska* (David M. Hopkins), 119
*Thermochemistry to petrology, Some examples of the application of* (Th. G. Sahama and D. R. Torgeson), 255
Thermodynamic stability, problems of, in magmatic theory for origin of quartzo-feldspathic massifs and veins (Hans Ramberg), 22
*Thermometry, granite, Liquid inclusions in* (Sturges W. Bailey), 304
Topography, of the Arctic Coastal Plain (Robert F. Black and William L. Barksdale), 106
*and sediments of the Arctic Basin* (K. O. Emery), 512
Torgeson, D. R., and Th. G. Sahama. *Some examples of the application of thermochemistry to petrology*, 255
Trask, Parker D. Review of: Soil mechanics in engineering practice. By Karl Terzaghi and Ralph B. Peck, 622
*Trivoli sandstone of Williamson County, Illinois* (Raymond Siever), 614
Troll, Carl. Strukturböden, Solifluktion und Frostklimate der Erde. Review by Ernst Antevs, 232
Tromp, S. W. The religion of the modern scientist (neo-materialism). Review by Kirtley F. Mather, 625
Tuttle, O. F. *Structural petrology of planes of liquid inclusions*, 331

*Umhlangakulu River, southern Natal, South Africa, Some remarkable wind-faceted pebbles from near the mouth of the* (Lester King), 413
*Universe, The abundance of the rare gases in the earth and in the* (Hans E. Suess), 600

Valle Grande (Ada Swineford), 308
Vanderwilt, John W., supervisor. Mineral resources of Colorado. Review by A. F. Hagner, 439
*Veins, quartzo-feldspathic massifs and, The facies classification of rocks: a clue to the origin of* (Hans Ramberg), 18
*Ventifacts, Pleistocene, east of the Big Horn Mountains, Wyoming* (Robert P. Sharp), 175
Verhoogen, Jean. *Radial diffusion and chemical stability in the gravitational field: a discussion*, 93
Victoria (C. A. Cotton), 289
*Villafranchian stratigraphy in southern and southwestern Europe* (Hallam L. Movius, Jr.), 380
*Volcanic ash, Great Plains Pleistocene, Source area of* (Ada Swineford), 307
*Volume and weight, Replacement and the equating of* (John Ridge), 522

Wahrhaftig, Clyde. *The frost-moved rubbles of Jumbo Dome and their significance in the Pleistocene chronology of Alaska*, 216
Weatherby, C. A., H. W. Camp, and H. W. Rickett, comps. International rules of botanical nomenclature. Review by E. C. Olson, 442
Wedge structures, description and origin (J. P. Schafer), 165, 169
*Weight, Replacement and the equating of volume and* (John Ridge), 522
Western Australia (C. A. Cotton), 294

Williams, Howel. The ancient volcanoes of Oregon. Review by F. M. Byers, Jr., 324

*Williamson County, Illinois, Trivoli sandstone of* (Raymond Siever), 614

Wind action. *Pleistocenes ventifacts east of the Big Horn Mountains, Wyoming* (Robert P. Sharp), 175

*Wind-faceted pebbles, Some remarkable, from near the mouth of the Umhlangakulu River, southern Natal, South Africa* (Lester King), 413

*Wind River Mountains, Wyoming, Stone nets, stone stripes, and soil stripes in the* (Gerald M. Richmond), 143

*Wisconsin, southern, Periglacial features in the Driftless Area of* (H. T. U. Smith), 196

Woo, Ching Chang. Review of: Acta Geologica Taiwanica, vol. 1, no. 1, 437

Review of: Bulletin of the Oceanographical Institute of Taiwan, nos. 1–3, 436

*Wyoming, Pleistocene ventifacts east of the Big Horn Mountains* (Robert P. Sharp), 175

*Stone nets, stone stripes, and soil stripes in the Wind River Mountains* (Gerald M. Richmond), 143

Yamasaki, Kazuo. *Geochemical researches in Japan*, 414

Zobel, O. J., L. R. Ingersoll, and A. C. Ingersoll. Heat conduction, with engineering and geological applications. Review by T. S. Lovering, 327

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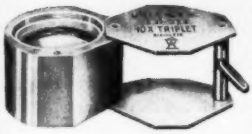

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**CONTENTS**

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|---|---|---|
| SECTION I. | General and Historical Papers | 3 |
| SECTION II. | Salt Dome Case Histories—Texas, Louisiana and Mississippi | 21 |
| SECTION III. | Mid-Continent Case Histories—Arkansas, Illinois, Oklahoma and Texas | 17 |
| SECTION IV. | Rocky Mountain Case Histories | 4 |
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## CONTENTS


| | | Page |
|---|---|---|
| FOREWORD | By John T. Galey | 1395 |
| SUMMARY OF MIDDLE ORDOVICIAN BORDERING ALLEGHENY SYNCLINORIUM | By Marshall Kay | 1397 |
| CAMBRIAN AND ORDOVICIAN ROCKS IN MICHIGAN BASIN AND ADJOINING AREAS | By George V. Cohee | 1417 |
| NEW YORK SUBSURFACE GEOLOGY | By E. T. Heck | 1449 |
| SUBSURFACE TRENTON AND SUB-TRENTON ROCKS IN OHIO, NEW YORK, PENNSYLVANIA, AND WEST VIRGINIA | By Charles R. Fettke | 1457 |
| TRENTON AND SUB-TRENTON OUTCROP AREAS IN NEW YORK, PENNSYLVANIA, AND MARYLAND | By Frank M. Swartz | 1493 |
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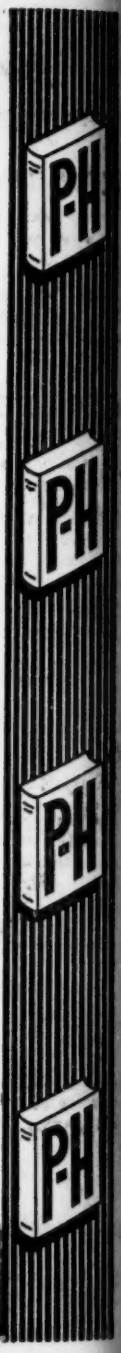


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